STRUCTURAL GEOLOGY

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STRUCTURAL GEOLOGY

CHAPTER I

INTRODUCTION

The central feature of *structural geology* as treated in this book is the interpretation of rock structures caused by earth movements. The outer limits of the subject are not clearly defined, for in one way or another it is interrelated with all other phases of geologic science, and with other sciences such as physics and astronomy. By some writers structural geology is broadly interpreted to include the study of primary structures caused by sedimentation and vulcanism, of topographic forms, of ore bodies, and in fact of all structures and forms represented in the architecture of the earth. The subject then becomes almost as broad as geology itself. In this volume these topics are referred to only incidentally for the reason that their really adequate consideration would interfere with the orderly presentation of the problem of rock deformation, which by itself makes a satisfactory unit for discussion. It is assumed that the student taking up structural geology will have obtained some knowledge of primary structures and of rock forms due to causes other than deformation in his various elementary courses.

The purpose of structural geology is the study and interpretation of rock structures, not for themselves, but for the light they may throw on stratigraphic problems, on economic geology, on the causes underlying the general configuration of the earth, and on the earth's history.

As a foundation for ordinary human activities it is but natural that the lithosphere or solid earth should be a popular symbol of strength and permanence; but the geologist sees abundant evidences that it has fared badly in the contest with environmental forces, past and present. It has been weak and incompetent; it has bent, crumpled, broken and mashed; structurally it has failed; in considerable part it now consists of structural ruins. That its failure, as disclosed by earth movement, is still continuing is abundantly shown by current changes in water levels and beaches, by landslides, and by earthquake disturbances.

The problem of the structural geologist includes the restoration of the ruins and a determination of the conditions and causes of failure. His problem is not rendered easier by the fact that it is seldom possible to see the structures in three dimensions, and that he must base his restoration on fragments of evidence seen at the surface or the very limited outlook of underground openings, or on inferences from environmental conditions. Furthermore, the geologist seldom sees rock failure in actual progress. If he does he may not recognize it because of the slowness of the movement. He arrives after the disturbance is over, and must infer the nature of the forces and processes from the results. In attempting to picture conditions in the inaccessible deep zones, he must make long-range inferences from the few available facts.

The study of structural geology naturally begins with the mapping and description of separate structures such as folds, faults, joints, and cleavage. Too often this has been regarded as the end and not as a step toward the understanding of the structural conditions as a whole. The necessity of integrating evidence and information from scant observations requires an understanding of the interrelations of structures, and of great group characteristics of a given environment or of a given kind of rock.

The use of the previous edition of this book as a text has disclosed certain shortcomings which, we hope, are at least in

part eliminated in the present edition, — which is almost entirely rewritten and largely extended. For instance, the field interpretation of cleavage seems to be difficultly understandable by students not having the advantage of instructional aid by teachers experienced in this phase of the subject. Attempt is made in the present edition to present this subject more fully, in the hope that this important aid in field work will be more largely utilized. Also, the old edition, in common with most textbooks in structural geology, was mainly confined to the deformation of hard rocks, at least by implication. In the present edition relatively more emphasis is placed on the deformation of soft and unconsolidated rocks.

The most important change in the new edition is in the general approach to the subject. The first part is devoted to a somewhat extended description of structures in the zone of observation. Later chapters deal with causes of deformation, general considerations, and with the unseen zone below. By this method of treatment we avoid complicating our discussion of the zone of observation with speculative considerations.

In order to preserve perspective and continuity the attempt is made to minimize definitions, descriptive details, and minute subdivisions of topics. This procedure is suggested by the writer's teaching experience, during which he has found that the presentation of the subject in short, more or less unconnected paragraphs, relating mainly to definition and detail, although it may give the student a certain empirical facility in field work and writing, fails to give him that general perspective of the subject which is so vital to an intelligent understanding of complex field conditions.

In the preparation of this text the writer is indebted to Professor W. J. Mead for valuable criticisms and suggestions arising from his own extensive field and experimental work; to Professor Edward Steidtmann for a summary of experimental results presented in the appendix; to Mr. Julian D. Conover for help in revision of the text; and to many other associates in structural field work for valuable suggestions.

While there are many references to published works, no attempt has been made to make the list exhaustive. The purpose has been rather to avoid the detail and complication inherent in an adequate summary of the work of others. There is, therefore, large obligation to many unnamed investigators.

CHAPTER II

A GENERAL SURVEY OF STRUCTURAL FAILURE IN THE ZONE OF OBSERVATION

We may direct our attention first to the structural failure of rocks within a comparatively few miles of the earth's surface. The characteristics of this region are disclosed to us by deformed rocks, some of which were once much farther below the surface than at present, but which have been brought within our range of observation by the erosion of overlying rocks and by numerous mine-openings. This may be conveniently referred to as our zone of observation.

Heterogeneous Nature of Movement

In this zone, some of the rocks have been deformed by rock flowage and some by rock fracture, both kinds of deformation being accompanied by folding, tilting, mountain-building, uplift and depression. The characteristic result of rock flowage is the production of laminated or foliated structures in rocks called slates, schists, and gneisses, but some rocks which have flowed do not show these structures, nor in fact any internal evidence of flowage. The results of rock fracture are described as joints, faults, fracture cleavage, and brecciation.

By rock flowage we mean "solid," "plastic," "massive," or "viscous" movement under pressure, either in hard or in soft, unconsolidated rocks. Flowage in molten masses is usually treated separately as an igneous phenomenon, — not as rock failure. No one of the descriptive terms used just above may be technically accurate and comprehensive, but the movement partakes of the characters expressed by all of them. The movement is not necessarily slow and continuous; in some cases there is geologic evidence that it is periodic. Rock

flowage in coherent rocks is characterized essentially by the parallel dimensional arrangement of the minerals, such as mica and hornblende, developed by recrystallization during the process. These minerals are present abundantly after the process, not before. Rock flowage is intimately associated with rock fracture, — the minute granulation and slicing of mineral particles, and even larger fractures, especially of the shearing type, being often included under the head of flowage. While rock flowage and rock fracture constitute two distinct types of deformation, there is almost complete gradation between the two, and much deformation is not accurately described by either term. A displacement may take place along a clean fracture, or along a fracture on which there has been local rock flowage, or along a zone of closely spaced parallel fractures with rock flowage affecting all of the intervening masses, or along a zone of rock flowage in which evidences of fracture planes are indistinct or altogether lacking. A single shear zone may show all of these features. In a large way a considerable zone of flowage may often be interpreted, in its relations to displacement and stresses, in much the same manner as a fracture plane.

In short, within our zone of observation, it is clear that rocks are heterogeneous in their composition and in their behavior under stresses.

Causes of Movement

Rock failure is evidence of overpowering stresses, but the causes and directions of these stresses are not so clear. Failure on a mountainous or continental scale points to great earth stresses of the kinds which have been variously ascribed to adjustments under gravity between earth masses of differing densities and elevations (isostatic adjustment), to adjustments under gravity of a solid shell to a shrinking centrosphere, a conception based on the supposed transfer of heat and magmas from the centrosphere outward, to tidal strains, to changing centrifugal pressures caused by changes in rate

of the earth's rotation, to igneous intrusion, or to combinations of these causes.

So clear is the evidence that great earth forces of this kind have been operative that other causes of movement have been perhaps underestimated or ignored in explaining local failure. Such are the pressures and changes of temperature attending the extrusion and intrusion of igneous rocks, in the vicinity of which there is often clear evidence of local failure, the recrystallization of rocks during long periods causing local changes of volume, the leaching out of substances near the surface causing voids and weakness and consequent slump under gravity, and other volume changes during weathering. When rocks are in a soft and incoherent condition, they are especially susceptible to local stresses. Mud, marl, sand and salt deposits crumble and slip as the deposits are slowly built up, either under air or water. Local loading by water and ice or rock materials may cause them to fail. Unconsolidated glacial deposits show a variety of joints, faults, and folds. In the settling, consolidation, and desiccation of any of these soft deposits, stresses are set up which result in local failure. When the deposits are seen later as hard rocks, it is difficult to determine to what extent the failures should be attributed to these early and local causes, acting during the soft, formative stages, and to what extent they are the result of regional deformation after the rocks have become strong and hard.

The part played by the forces of crystallization in initiating earth stresses is yet but little understood. Growing crystals have been found experimentally to exert considerable linear forces, and there seems to be evidence in rocks that these forces have been sufficient to widen openings or to expand the rock mass. Crystallization may also contract the rock mass. The manner in which crystal habit asserts and maintains itself, even under the most intense conditions of metamorphism, is one of the most impressive facts of geologic observation. It is the custom usually to explain such facts on the basis of adaptation to environment, and to put the emphasis largely

on the environmental conditions as determining the outcome. In the development of crystals, however, it is clear that these conditions have not been sufficiently intense to interfere with or overcome the tendency of the crystals to take whatever form best suits their atomic structure — in other words, to develop their own habit. The philosophy of the precise relations between inherent crystallizing power and environmental forces is not understood; but enough is known to warrant the suspicion that the cumulative effects of forces of crystallization may themselves initiate earth stresses of a high order of magnitude.

Better criteria are needed for the discrimination of rock structures due to local stresses of the kinds indicated, from the results of failure under the greater regional earth stresses. Of course, there is no clean-cut separation between the two. An accumulation of minor and local causes may give rise to relatively large earth movements, and conversely major earth movements are frequently resolved into a complex of minor related structural phenomena.

Angular Relations of Rock Structures to Causal Stresses

Just as rock structures in themselves do not indicate all the causes of failure, neither do they indicate clearly the directions of application of stress. On the whole the geologist's attempt to relate specific structures with specific stress systems has not been highly successful. The various structures resulting from rock failure have usually been explained on the simple conception of the application of a non-rotational stress — either tension, causing elongation in the direction of pull, or simple compression, producing a shortening parallel to the principal stress and elongation at right angles to it. A fold, for instance, is assumed to indicate application of stress normal to its axial plane; a set of compression joints is taken to indicate application of stress at 45° to the fractures; cleavage is taken to indicate application of pressure normal to its plane. Experimental work on rock deformation has been con-

ducted mainly with the same limited assumptions, and the results have been widely quoted and applied to the interpretation of rock structures in the field. These conceptions may be correct in so far as the immediate features are concerned; but the stresses thus assumed may be only minor components of the major causal stress and give no clue to its direction.

Much less attention has been paid to the conception that compressional forces may be rotational, that is, that they may be applied in the form of couples. Under this conception, the net result is a shearing between the heterogeneous rock units along planes ranging from parallel to 45° to the principal axis of stress, the shearing usually accompanied by local tension — it being conceived that no matter what the origin of compressive stresses and their angle of application, when applied to the heterogeneous rock masses constituting the earth they tend as a whole to act in couples, and to be resolved into components acting in directions inclined to the resulting planes of movement. A mountain-making movement under this conception is a shear of certain rock masses over others, resulting in faults, joints, folds, and cleavage. Tensile stresses may be minor consequences of such a shear. Accumulating field observations favor this view of the dominance of shear. It is the view also which geologists have commonly applied to an assumed shear of a thin brittle crust over a thin mobile zone below, though curiously enough not to the local structures that can be observed.

In later chapters, dealing with fractures, rock flowage, and folds, evidences of shear are cited.

Distribution of Movements

Structural failure within our zone of observation, whether by fracture or flow, has not been confined to any particular plane or formation, but is so distributed as to indicate that adjustment of rock masses under deforming stresses has been accomplished by movements in many planes and zones, in many formations, in all directions, and with all inclinations. Rocks

in this zone as a whole have not yielded to stresses as homogeneous masses. In fact, even down to comparatively small units of volume the rule has been heterogeneity. No matter how homogeneous the formation may seem, rock movement has disclosed zones of inherent weakness along which the movement has been largely concentrated.

Within our zone of observation, it is difficult to say inductively whether or not there has been more movement or less movement with depth. Neither is it possible with any satisfactory degree of definiteness to discern controlling attitude or pattern in the complex of movement zones. The zones range from vertical to horizontal, are parallel or intersect. The original horizontal position of stratified rocks naturally suggests dominance of the horizontal element in movements affecting them, because of resolution and transmission of forces along bedding planes; but the beds soon become inclined or vertical when deformed, and disturbed zones may be anything but horizontal. The less deformed masses between may have almost any shape. Locally they may be discoidal, or sheet-like, or oval, or rod-shaped, or rhomboidal. Interesting attempts have been made to discern some controlling pattern, both in large and in small structural features, but subjective hypotheses enter to so large an extent that the reality of the pattern presented is often not convincing to others.

Possible Increase of Rock Flowage with Depth

Within a few hundred or at most a few thousand feet of the surface, fracturing, much of it open, is clearly the dominant process, although even here soft rocks may yield by flowage. In the lower part of the zone of observation combined fracture and flowage is the rule. Fractures here are more commonly of the closed, shearing type. It has been easy to assume that this combination of fracture and flowage is merely transitional to a *zone of flowage* below. The fact that rocks which have been deeply buried are often highly schistose as a result of rock flowage has been cited as indicating increased rock flowage

with depth. The writer has shared in this view. From some familiarity with ancient and formerly deeply buried terranes, he is not sure, however, but that a careful inductive study of field sections requires very considerable qualifications of this generalized conception. Many instances may be cited of rock flowage occurring high in the geologic section and rock fracture below. On the whole, the oldest rocks undoubtedly show the most abundant evidences of rock flowage, but even in these rocks such evidences are localized in relatively narrow and numerous zones. It should be remembered that these rocks have suffered more periods of deformation, some near the surface and some deep below, than the younger rocks. The evidences of flow which they present do not necessarily indicate that all the flowage occurred at great depths. It may be said that plutonic intrusions of great mass often cause rock flowage in the adjacent formations, and that so far as such intrusions are more numerous with depth, rock flowage may increase. On the other hand, some plutonic intrusions in younger series which have not been very deeply buried likewise cause rock flowage. It is also certain that shearing movements, resulting in displacements which we call faults, have extended down to the bottom of our zone of observation. These partake of the nature of rock fracture in their confinement to places and their relations to stresses, although whether the processes be called flow or fracture is partly a matter of definition.

In conclusion, both fracture and flowage are to be seen in all parts of the zone of observation. The existence of a zone of flowage in the unseen below is not yet proved (see Chapters VII and XIV).

Movements Still Continuing

In discussing rock movements we are apt to fall into the use of the past tense for the reason that in most cases only the results of movements are to be observed and the movements themselves antedate our observation. That movements are

now going on, however, is variously indicated by changes in water levels and shore lines, by earthquake disturbances, by rock slides, and other evidence.

There is abundant geologic evidence that movements in the past have not proceeded uniformly, but have been periodic. Periods of activity have been followed by what seem to be longer periods of quiescence. We do not know whether on the whole the present movement in the earth's crust is proceeding more or less rapidly than in the past, whether we are in a period of quiescence or activity. The slowness and localized distribution of recognized movements in historic time have led to a rather general assumption that this has been a relatively quiescent period. On the other hand, the cumulative effects of these slow movements have been considerable, and when surveyed in retrospect through the long vistas of future geologic time this might appear to be a part of a period of active disturbance. It is thought, for instance, that mountain building on an Alpine scale is now proceeding in the Dutch East Indies.

CHAPTER III

DEFINITION OF TERMS

Structural geology treats of the arrangement of rocks or the architecture of the earth's crust. As thus defined it includes the study of structures caused by sedimentation and vulcanism, and even by erosion. The usage of different writers has varied greatly in setting the limits of the field of structural geology. Practically all, however, emphasize structures due to earth movements, and this emphasis is followed in the present book. Structures other than those caused by rock movements are here treated only incidentally.

Tectonic or geotectonic geology has been used more or less synonymously with structural geology. The word tectonic is defined in the Century Dictionary as "of or pertaining to building or construction." On the whole, there is perhaps a little stronger tendency to restrict this term to rock deformation during earth movements than is apparent in the use of the term structural geology, but even its use is not strictly confined to secondary deformation. Geikie,[1] for instance, treats of tectonic mountains as including mountains of deformation and mountains of accumulation such as volcanic cones. In the present volume the term is confined to deformation.

Dynamic geology[2] treats generally of causes, agencies, and processes, but is frequently more or less confined to earth movements, and is thus used in much the same sense as structural or orogenic geology, or diastrophism.

Diastrophism, as defined by Chamberlin,[3] "includes all

[1] Geikie, James, Structural and Field Geology, D. Van Nostrand Co., New York, 4th ed., 1920, p. 402.

[2] Chamberlin, T. C., and Salisbury, R. D., Geology, vol. 1, Henry Holt and Co., 1904, p. 1.

[3] *Loc. cit.*, p. 2.

crustal movements whether slow or rapid, gentle or violent, slight or extensive." The emphasis is on movements and processes of deformation, but consideration of the structural results is of course included. *Megadiastrophism* relates to the larger earth movements, such as those forming continents and mountains.

Orogenic geology, orogeny, or orogenesis is a phase of diastrophism which treats of the movements involved in the formation of mountain ranges. As the movements and forces involved are of course much the same as those deforming rocks generally, the term has come to be used in some quarters more or less synonymously with rock deformation. In this book it is confined to mountain-making movements. *Epeirogenic geology* is a phase of diastrophism which treats of the movements of continents.

Deformation is the change in form of a rock under stress. A more technical term for the same change is *distortion.* In discussing deformation the emphasis is on results of movements or diastrophism, but consideration of the movements themselves is of course included.

Deformations produced by internal causes, such as chemical change, recrystallization, etc., have been called *endogenetic.* Deformations caused by external forces, such as those involved in tectonic disturbances, have been called *exogenetic.*

Change in volume is called *dilatation,* which is positive or negative according as the volume increases or decreases. Dilatation is seldom absent in deformation, and ordinarily the term *deformation* is used to include dilatation.

It is apparent from the above summary that some of the more general terms used in structural geology lack precision in definition and usage and that they are more or less overlapping. This is perhaps inevitable, because of the very nature of the subject and its vague limits. All of the terms, however, cover certain aspects of earth movements and their structural results. Probably more explicit definition would be desirable, but the old terms are all useful

and are so well entrenched in the literature that it will be a long time, if ever, before more precise usage becomes general.

Terms used in the description of rock movements themselves may be somewhat more exactly defined, though even here variability in usage is by no means lacking. The mechanics of rock movements constitute a highly technical subject, bristling with mathematical and physical difficulties. Few geologists have the necessary knowledge of mechanics to understand and interpret its applications to rocks, and unfortunately few mathematicians, physicists, or engineers have known enough about earth conditions to make effective application of the principles of mechanics. Some phases of the subject have as yet been barely touched, even by specialists. An elementary text of this kind is not the place for an adequate exposition of the mechanical aspects of the subject, even if the writer were qualified to make it. However, a few simple mechanical principles are necessary for the understanding and interpretation of rock structures, and these require a few elementary definitions in order to avoid clumsy circumlocution.

In geological usage the force tending to deform a rock is often spoken of as a *stress;* the deformation, that is, the change in shape or size, resulting from the application of the stress, is a *strain.* The main point to remember is that what we see in rock deformation is strain. We do not see, but must infer, stress. Very often in geological descriptions these two terms are confused.

More technically, when external forces are applied to a rock mass, tending to change its shape or size, there are internal forces between the particles of the mass which resist such deformation.

"The action and reaction between any two adjacent portions of a body constitute a *stress.* Considering any two such portions separated by a plane, the forces exerted between them may have any direction relative to that plane. It is convenient to consider

these forces as resolved into components normal[1] and tangential to the plane. The stress made up of the tangential components is called a *shearing stress*, and may have any direction in the plane considered. The normal stress is a *tension* if it resists a tendency of the two portions of the body to separate, and a *compression* if it resists a tendency of the two portions to approach each other.

"Whatever stress condition exists at any point of a body, there are always three rectangular planes upon each of which the resultant stress is normal. Three intersecting lines perpendicular to these three planes, respectively, are called the principal *axes of stress*.

"Upon any plane not perpendicular to a principal axis the stress has in general a tangential component. It may be proved that, for planes parallel to one principal axis, the tangential or shearing stress has the greatest intensity upon planes inclined 45° to the other two principal axes."[2]

In the directions represented by the three principal axes of stress, the stresses may all be of the same intensity; two may be of equal intensity and the third greater or less; or, most commonly, the three may all be different. In this last case it will be found that the stress parallel to one of the three axes is the greatest stress existing within the rock in any direction whatsoever, that the stress parallel to one of the other axes is the least stress in any direction, and that the stress parallel to the third axis has some intermediate value. The corresponding axes are accordingly called the *greatest, mean*, and *least axes of stress*. The difference between the greatest and least stresses is called the *stress difference*.

A rock has a certain amount of elasticity and under small stresses it will resume its original form when the deforming force is removed. Thus, when a block of rock "is subjected to normal stress in one direction, being unsupported in other directions, it is perfectly elastic until the intensity of normal stress reaches a certain limiting value, called the *elastic*

[1] *Normal*, throughout these discussions, is used in the sense of *perpendicular*.
[2] Hoskins, L. M., Flow and fracture of rocks as related to structure: 16th Ann. Rept., U. S. Geol. Survey, Pt. 1, 1896, p. 867.

limit."[1] When the stress exceeds this value, deformation is partly permanent, that is, the rock will not resume its original form when the force is removed. If the deforming force is increased the rock will finally rupture. The *breaking* or *ultimate strength* of the material is defined as the greatest intensity of normal stress existing before rupture. In the case where rocks are stressed in more than one direction, which is the usual case in structural geology, these definitions require slight modification; the elastic limit and the ultimate strength may then be said to be those values of the *stress difference* at which permanent deformation and rupture, respectively, will occur.

Repeated application of stresses weakens a rock. This loss of strength is called *fatigue.*

According to Hoskins a body is said to be *strained* when the relative positions of the particles undergo any change. The term strain is applied to changes either in the volume or the form of the body, therefore it includes both dilatation and distortion. Rotation or translation of the body as a whole may accompany strain but forms no part of it. Strain may be expressed in conventional terms of three principal strain axes, which are mutually perpendicular. It should be emphasized, however, that *strain* is simply a geometrical conception, expressing change in shape or volume, and does not imply knowledge of the stresses to which it is due. As will be seen below, the principal strain axes may or may not be parallel to the principal axes of stress.

In the directions represented by the three strain axes, the strains, or changes in length, may have the same or different values. When different, the direction in which the largest amount of shortening has occurred is called the *least axis,* the direction of the greatest elongation is called the *greatest axis,* and the third direction, perpendicular to the other two, is called the *mean axis, of strain.* In this case there is a change in shape and the strain is called *distortion.* If the

[1] *Loc. cit.,* p. 845.

strains along all three axes are equal, it signifies that there has been cubical contraction or expansion, that is, that the mass has contracted or expanded in the same ratio in all directions. This change of volume only is called *dilatation* (p. 14). Distortion and dilatation are ordinarily combined in rock deformation, but distortion is more obvious and more significant in most of the rock structures dealt with by the structural geologist.

Relations between Stress and Strain

Stresses may be so applied that the normal components are equal in all directions, or in other words, in such a manner that the stress difference is zero. There is then no change of form, but the volume change which we have called dilatation. If the stresses are compressive, this is sometimes called *uniform* or *hydrostatic pressure*.

Stresses may be applied unequally, producing distortion. This is sometimes called *non-uniform pressure*.

Non-uniform compressive stress may be so applied that the greatest axis of stress, sometimes called by the geologist the greatest pressure, remains continuously parallel to the least axis of strain or direction of most shortening. The result is a simple shortening of the rock mass in this direction, and elongation at right angles thereto. No matter how far the deformation is carried, the least, mean and greatest axes of strain remain parallel, respectively, to the greatest, mean, and least axes of stress. This is called a *pure* or *non-rotational* or *irrotational strain*. " Non-rotational " refers to the lack of angular change between, for example, the greatest axis of stress and the least axis of strain. It is the kind of stress-strain relation shown when a rubber ball is flattened on a table by vertical pressure, or when a cube of building stone is crushed by pressure normal to one of its faces. About the same elongation of the ball could be accomplished by tension parallel to the table.

Non-uniform stress may be so applied that the strained

body is continuously rotating under it. In this case the greatest axis of strain is continuously in angular relation to the least axis of strain. This is called a *rotational strain*, or *shear*, or *shearing strain*, or *scission*, or *tangential strain*, or *differential moving*, or *slipping*, or *gliding* (see Figs. 1, 3, and 6). It is the kind of stress-strain relationship shown when the cards of a pack lying on a table are slid over one another by pressure parallel to the table-top. The action of the applied forces is that of a couple, or is similar to that of a pair of shears, from which analogy the name is derived. Within the body these forces are so resolved that there are both normal and shearing

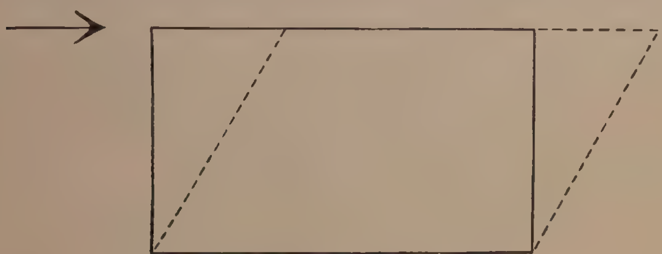

FIG. 1. Illustrating simple shearing.

components (p. 21), — the greatest normal components, or principal axes of stress, being inclined to the applied shearing forces. The resulting shortening and elongation may be the same as those produced by irrotational movement, and are likewise described by reference to the principal axes of strain; in fact, the net result might have been accomplished equally well by a non-rotational strain combined with rotation of the body as a whole. In rotational strain, however, the stress-strain relationships may be quite different from those in the non-rotational strain; thus, while at any instant the greatest tendency toward shortening is in the direction of the greatest normal stress, the accompanying rotation may prevent the continuation of the same relation of the stress and strain axes, and the axis of the greatest *total* shortening may at any instant be at a considerable angle to the greatest normal

stress and the axis of shortening then occurring. The direction of shortening is always inclined to the applied shearing stress. What we see in rocks is elongation and shortening, and usually we cannot tell whether the stress producing this result has been rotational or non-rotational in its application.

Planes of Maximum Shear

In any deformation, whether rotational or non-rotational, tensional or compressional, excepting only the case where the stresses are equal in all directions, there are, as stated on page 21, components of the stress which act parallel or tangential to most planes within the mass. The surfaces or planes along which this stress is at a maximum are called planes of maximum shear, and are the planes along which the rock has the greatest tendency to break by shearing. In a non-rotational strain the planes of maximum shear are theoretically 45° to the direction of greatest pressure, which is also the axis of most shortening or least axis of strain; actually, because of internal friction (see p. 35), they are usually at smaller angles to this direction. In a pure rotational strain the planes are again about 45° to the shortening, or least axis of strain, but the position of the more important one is nearly parallel to the greatest applied pressure (or shearing force) while the other is at 90° or steeply inclined to it (Figs. 3 and 6). There are also all combinations of the two cases, giving intermediate positions for the planes of maximum shear.

The use of the term *shear* or *shearing planes* is a source of much confusion. The geologist partly inherits this confusion from the miscellaneous usage of the term in mechanics, but his own usage has not helped to clarify its meaning. Application of external forces to a body causing a rotational, or differential sliding movement, as in the slipping of one sedimentary bed over another, or the slipping along a great thrust fault plane, is called shearing, meaning that the deforming stresses are tangential to the plane of movement and that the mass as

a whole is subjected to forces acting as a couple. But within the mass, the external forces accomplishing this deformation are so resolved that on nearly all possible planes there are both normal and tangential or shearing stresses, — varying in intensity according to the angular position of the plane. Likewise in simple elongation or shortening under non-rotational tension or compression, within the mass there are on all planes, except those perpendicular to the principal stress axes, tangential components of the stresses; there are shearing planes, and planes of maximum shear. The rock mass may or may not yield along these shearing planes, but they are there. In short, shearing planes and stresses, which may or may not be expressed by actual failure, are present whether the stresses acting on the mass as a whole are applied non-rotationally as simple tension or compression, or rotationally as shear. As expressed by Hoskins: [1]

> "Only under exceptional conditions can there be strain without the occurrence of sliding along certain planes, or stress without the existence of tangential stress on certain planes. The only case in which shearing stresses are absent is that in which the normal stress has the same intensity on all planes, whatever their directions. The only case in which there is no sliding or tangential strain is that of a simple dilatation — the elongations or the shortenings in all directions being equal."

The Strain Ellipsoid

It is sometimes convenient to describe strain and the relationships between strain and stress in terms of a conventional form known as the *strain ellipsoid*. (See Figs. 2 and 3.) Any imaginable sphere of a solid body when deformed in a single homogeneous [2] strain becomes a strain ellipsoid. This ellipsoid has three principal and mutually perpendicular axes — the greatest, mean, and least, which correspond to the three prin-

[1] *Loc. cit.*, p. 868.

[2] By Hoskins a strain is said to be *homogeneous* "when any two portions of the body which are similar in form and similarly oriented before the strain are also similar in form and similarly oriented after the strain."

cipal axes of strain. Where we know the directions of elongation and shortening in a rock, we can conceive of these in terms of the strain ellipsoid, and thus visualize the deformation which the rock has undergone. It cannot be too strongly emphasized that the strain ellipsoid in itself does not show the directions of application of the forces. It indicates merely the result of deformation. The forces may have been applied as tension or compression without rotation, causing simple elongation and shortening, or with rotation as in shear. One

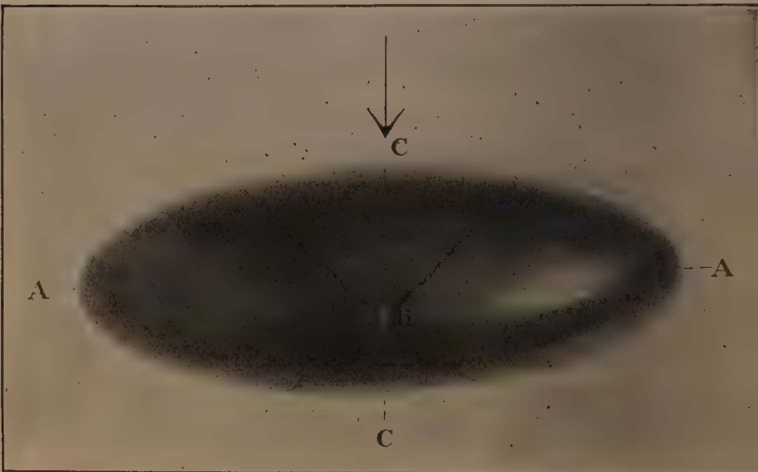

FIG. 2. Strain ellipsoid resulting from deformation of a sphere by compression with non-rotational strain. The three principal stresses are unequal, and are continuously parallel to the three principal axes of strain AA, BB, CC. The black lines indicate the planes of no distortion or planes of maximum shear. It may be noted that the same ellipsoid could be formed by tension, with resultant elongation along the axis AA and unequal shortening along BB and CC.

of the most common mistakes in geological field work is to assume that the flattening has been normal to the greatest pressure, or greatest axis of stress, that is, that the strain has been non-rotational. In a few cases it may be possible to determine from other field evidence that this is true, but in the

great preponderance of cases the question must be left open whether the result was brought about in this manner or by shear.

The strain ellipsoid affords a ready empirical method of determining the approximate location of the planes of maximum shear in either rotational or non-rotational strains, without the necessity of detailed mathematical calculations. In a strain ellipsoid with three unequal principal axes there are only two cross-sections which are circular in outline (see

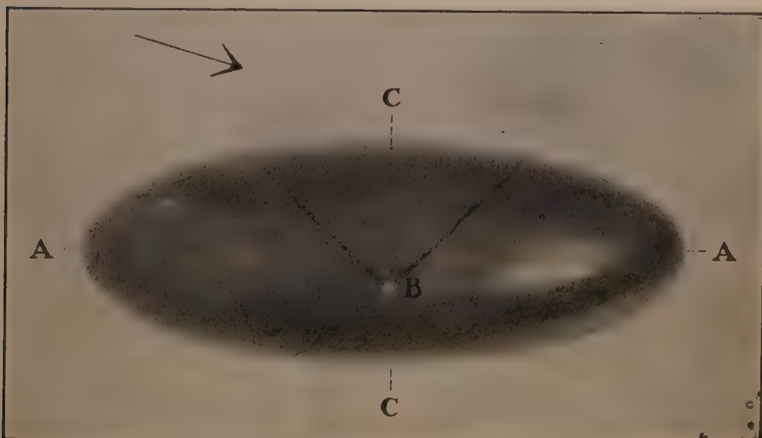

FIG. 3. Strain ellipsoid resulting from deformation of a sphere in rotational strain, — i.e., by shear. The three principal stresses are unequal, are inclined to the applied shearing stresses, and are continuously changing in direction. The strain axes are likewise inclined to and constantly rotate relative to the applied shearing stresses; the axes of total elongation and shortening, AA and CC, may differ from the axes along which the greatest tendency to elongation or shortening exists at any given instant. The black lines indicate the planes of no distortion or planes of maximum shear. It will be noted that the strain ellipsoid has the same shape as that formed in a non-rotational strain in Fig. 2, but that its angular relations to stresses are different.

Fig. 2). If the student will take a model of a strain ellipsoid in hand he will soon convince himself that this is the case. These planes, which are called *planes of no distortion*

because they preserve a circular cross-section similar to a section of the original sphere, are also the planes of maximum shear.

Given, then, a rock structure in which the directions of elongation and shortening, or in other words, the position of

FIG. 4. Wire netting model undeformed. See also Figs. 5 and 6.

the strain ellipsoid, is known, it is possible to locate the planes of maximum shear, without regard to whether the stresses have been tensional or compressional, rotational or non-rotational. As to the latter, usually we cannot tell. Application

of this principle is made in later chapters in discussing structures due to rock fracture.

A simple device for two-dimensional illustration of the position of strain ellipsoid and shearing planes in both rotational and non-rotational strain is shown in Figures 4, 5, and 6. A cardboard upon which is inscribed a circle is laid between two sheets of wire netting. The three are then fastened together by a rivet in the center of the circle. A wooden hinged frame fastened to the netting allows and controls the distortion of the netting, while the interior sheet remains undistorted. A

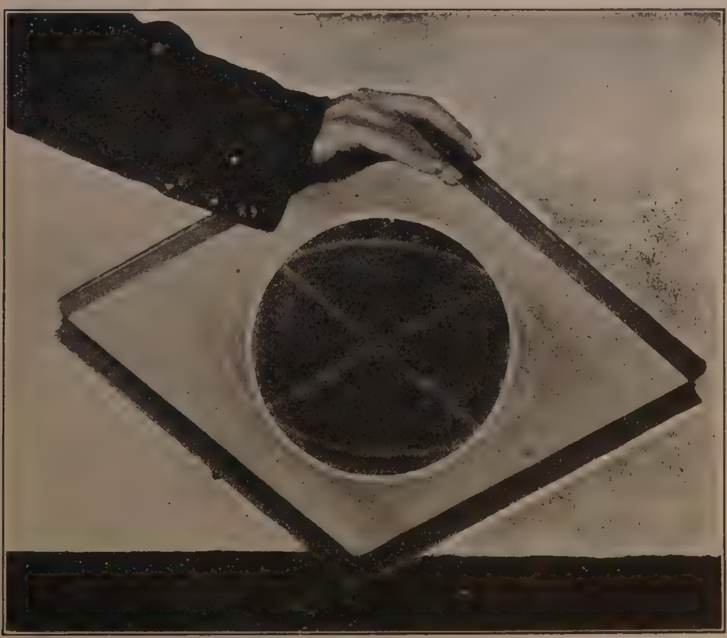

FIG. 5. Wire netting model deformed in non-rotational strain. Straight lines connecting intersections of circle and ellipse mark positions of "planes of no distortion" or planes of maximum shear.

circle and diameters are painted on the netting corresponding with those on the central sheet. When the screen is distorted the circle on the wire becomes an ellipse, or a cross-section

through the greatest and least axes of a "strain ellipsoid," which is superposed upon the undeformed circle of the cardboard.

In Figure 5 a non-rotational strain is represented, called "pure shortening and elongation." The circle elongates normal to the pressure. The planes of no distortion, which are the planes of maximum shear, stand normal to the surface of the screen. Their intersections with the plane of the screen are to be seen at about 45° to the pressure. It will be noted

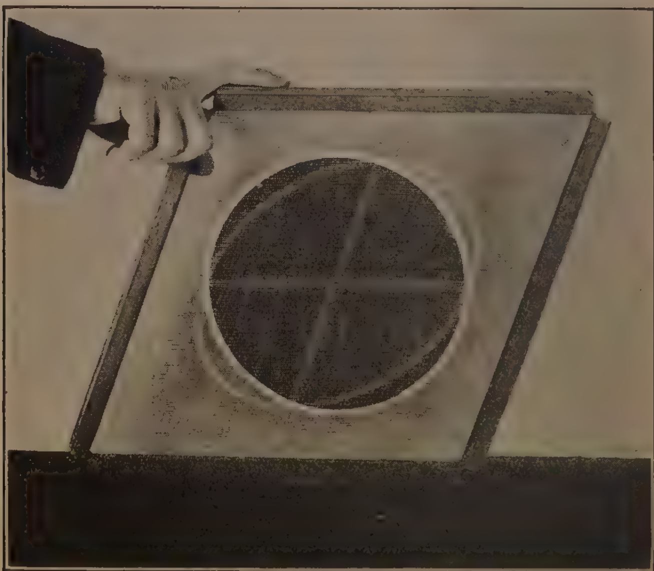

Fig. 6. Wire netting model deformed in rotational strain, or shear. Straight lines connecting intersections of circle and ellipse mark positions of "planes of no distortion" or planes of maximum shear.

that the lines representing the planes of shear are parallel to the wires. The distortion of the screen actually occurs by shearing of the wire mesh. This should make clear the fact that the painted lines of "no distortion" represent actual shearing planes.

In Figure 6 the strain is a rotational one. A strain ellipse

is produced by shearing of the top over the bottom of the model, obviously by movement along the shearing planes of the wire mesh. The planes of no distortion are indicated as before. It will be noted that they have the same relations to the ellipse as before, though the pressure has been applied at a different and varying angle. It is evident that the net result is the same as in non-rotational strain, so far as the shape of the strain ellipse is concerned. The rotation of the one figure in space would make it coincide with the other.

The model of course shows only the cross-section of the ellipsoid through the plane of the greatest and least axes of strain. It does not show the position of the mean axis. Great care must be taken, in this as in other two-dimensional representations of deformation, to keep clearly in mind the relation between the cross-section and the third dimension.

Summary

When rocks yield to stress, it is by undergoing some combination of dilatation, simple elongation and shortening, and shear. In all cases, except dilatation, there is shearing within the mass. Torsion and twisting and bending are only special cases involving various combinations of these simpler strains. Dilatation in rock deformation is not easily apparent, and not much is known about it. Structural geology deals mainly with deformation under non-uniform stresses and therefore with changes of form. These deformations, or strains, may frequently be discerned with considerable accuracy, but their relations to various possible combinations of causal stresses are usually not so clear.

The definitions and mechanical principles outlined above apply to all solid bodies, including rocks. Whether the rock be deformed by fracture or flow, or by some combination of the two, it is possible to describe the deformation in these terms. It must also be remembered, however, that a rock is not an ideally homogeneous substance, but has its own peculiar properties which sometimes prevent it from registering clearly

all the stresses acting upon it. Thus, while theoretically it tends under compression to fracture along planes of greatest shear, actually, because of initial texture or heterogeneity, the resulting structures may depart more or less from the theoretical positions.

These general considerations are applied specifically to individual rock structures in the following chapters.

CHAPTER IV

JOINTS

General Description

Joints are rock fractures or fissures along which there has been little or no movement. A joint that is tight and inconspicuous is sometimes called a *slip* or *rift*, or *blind joint*. The term *rift* includes also incipient joints, or surfaces along which the rock is not actually broken but merely strained and weakened. Where closely spaced tight joints give the rock a

Fig. 7. Jointing in granite, Cape Ann, Mass.

capacity to part along parallel surfaces, the structure may be designated as *fracture cleavage* (see pp. 148–158). Where there has been considerable movement along a joint surface the structure is usually called a *fault*. If the movement is slight and inconspicuous, it may still be called a joint; there is no sharp demarcation between the two structures. Where filled with later infiltrated mineral matter the joint or fault may also

be called a vein, and if the filling mineral is commercially valuable, the vein may be called a *lode.* Whether in such cases the structure is to be called a joint or a vein or lode depends on whether the structural feature or the filling is to be emphasized.

Fig. 8. Vertical joint planes in intersecting sets in a horizontal sedimentary formation. After Kindle, U. S. Geol. Survey.

Joints are usually "near plane" surfaces, but may also be curved or warped. The more common attitude is vertical,

or nearly so, but there are also many inclined and horizontal joints.

Joints are common to hard and soft rocks, to surface and deep-seated zones. Their occurrence in unconsolidated surface materials is especially discussed in Chapter IX.

Joints are more abundant near the surface than deep below, due to the fact that surface rocks have opportunity to expand under the forces of weathering there existing. Under surface conditions, also, joints tend to be widened and become conspicuous, even though they continue below as tight and less apparent structures.

Joints are so all-pervasive that they are often not recorded in geologic mapping; their very number and diversity of attitude may make it impracticable to map them except in very detailed surveys. Nevertheless, detailed observation is often desirable for the light it may throw on the manner of deformation of the rock, or of the region as a whole. The restriction of joints, or of certain types of joints, to certain formations or areas may be correlated with some distinctive feature of the geologic history. Their relations to topography and drainage are often significant, because they afford planes of weakness for erosion. Where filled with vein matter of economic importance, the necessity of careful observation of their attitudes and structural relations is of course obvious. In rock quarries, the number and distribution of joints may determine also whether the rock may be extracted in blocks of suitable sizes and shapes for building or other purposes. Still further, joints have much to do with determining the supporting strength of rocks encountered in mining operations, and in large construction projects such as canals or dams.

Causes of Jointing

Joints are caused by earth stresses attending general crustal movements such as are discussed in Chapter IX, and by expansion and contraction of rocks due to temperature changes, mineral alterations, and variations in moisture con-

tent. Particularly to be noted are the volume changes in the cooling and contraction of igneous masses, in the drying, settling, and contraction of sediments, and in all rocks under conditions of weathering. Joints formed as the result of volume changes are sometimes classified as expansion joints and contraction joints. Expansion joints are those formed during the expansion of a rock; the stress at the point of rupture is compressional. Contraction joints are those formed by the contraction of the rock; the stress at the point of rupture is tensional. Earthquake shocks also may be the immediate cause of jointing, though it is suspected that these act only as the triggers which set off stresses already existing.

From the moment any rock is formed, whether igneous or sedimentary, it is undergoing successive cubical expansion and contraction, and these continue during all phases of the metamorphic cycles through which the rock passes. These changes alternate faster at the surface than farther down, and hence the greater abundance of joints near the surface. However, of the rocks which have been deeply buried, there are probably few which reach the surface without preëxisting joints or planes of weakness; and the jointing which can be definitely related to surface alterations tends to follow these earlier structures, making it difficult to discriminate between the two. The tendency has been to ascribe the deforming stresses to general earth forces acting from without, especially in the case of the long continuous joints, sometimes called "master joints," which cut across rocks more or less regardless of type. A clearer understanding, however, of the part played by volume changes in all the various phases of the development and metamorphism of rocks, is leading to an appreciation of the importance of this factor in the formation of joints.

The causes of jointing may sometimes be inferred from the field relations. Thus there may be joints in recently cooled lavas which obviously have not been subjected to exterior forces. These may be irregularly distributed, or in

patterns such as basaltic parting or a general radial or concentric arrangement. They are clearly related to the contraction of the crystallizing and cooling mass, together with settling, in some cases, due to removal of support at the time of extrusion (see p. 73). Similarly joints may be very abundant in flat-lying, partially consolidated beds of sediments, which plainly have not been disturbed by great exterior forces. One of the causes in this case is doubtless the change in volume incidental to the drying and settling of the beds. Mud cracks are one manifestation of this process. Joints formed in this manner are likely to be limited to particular beds and may die out above or below; there may be evidence that jointing in a given bed was complete before the next layer of sediment was deposited. They are likely to be especially abundant near the contacts of different beds or formations (a fact often noted by well-drillers in search of water). Joints of this general type stand in various directions and attitudes. They may intersect, but more commonly one joint either dies out before reaching another, or is cut off by another joint crossing it at an angle.

The processes of weathering are highly favorable to the development of joints, as may be seen in almost any quarry section or natural exposure. Near the weathered surface the rock is likely to show many joints, more or less open, while below this surface the joints are usually less numerous and less conspicuous. Obviously temperature changes, frost action, the roots of vegetation, and volume changes due to chemical and mineral alterations, are active in breaking the surface rock. In the end-product of weathering the rock is so thoroughly disintegrated that joints, as entities, disappear in a confused mass of broken and altered weathered material.

Local changes in the loading of a rock brought about by erosion and deposition may change pressure conditions sufficiently to cause joints of small extent.

Attitude of Joints in Relation to Causal Stresses

An understanding of the attitude of joints with reference to the deforming stresses is sometimes useful in analyzing the structure of a deformed region, and especially in predicting unseen extensions of joints or veins and in the interpretation of folds.

Joints are caused by tension or by compression, or by some combination of these stresses as in torsion or cross-bending.

Fig. 9. Results of crushing wooden blocks by non-rotational stress. Note tendency of fractures to follow shearing planes 45° to the pressure (which was from above) regardless of the grain of the wood.

Under tension the joint plane is usually normal to the pull. Inclined shearing stresses are present in tension just as they are in compression; but ordinarily these do not find expression because rocks, with their low tensile strength, break more easily by tension than by shear along inclined planes. Exceptionally, however, the break will follow the shear planes.

Joints formed by compression tend to follow planes of maximum shear inclined to the principal stress. The lengthening of the mass normal to the direction of compression may locally cause tension joints transverse to the lengthening.

The angular relation of the planes of maximum shear to the axis of applied stress may vary from 90° to 0, depending on whether the compression is of a rotational or non-rotational nature, or some combination of the two.

Simple non-rotational pressure (see p. 18) is the kind ordinarily used in laboratory experiments and is the kind most frequently assumed by geologists in discussing fractures in

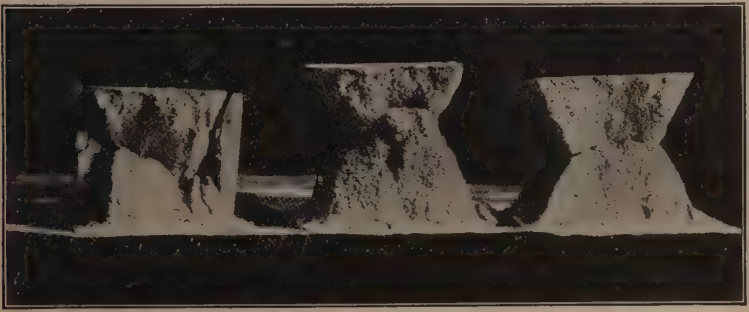

FIG. 10. Fracture of building stone (brown sandstone) along shearing planes. After Buckley.

rocks. Under such a pressure the rock breaks along one or more planes, at angles of somewhat less than 45° to the principal axis of stress. Why the planes should be less than 45° to the principal axis of stress is not entirely clear, but is explained by the phenomenon of internal friction.[1] It is an empirical fact finally determined by the inherent qualities of the rock. If the rock is free to escape on all sides at right angles to the principal stress, it may break along many intersecting shearing planes. The only limitation is that these planes must ordinarily be at angles of 45° or less to the axis

[1] For brief discussion of this still imperfectly understood subject see Merriman, Mansfield, Mechanics of materials: Wiley and Sons, 1916, pp. 375–382.

of principal stress. If the rock is free to escape only on one side the breaking occurs only in shearing planes inclined toward the free side.

It is probable that rotational strains are fully as common in the earth as those of the non-rotational type. In a strain having a rotational element the planes of maximum shear may vary from parallel to 90° to the applied stress (see p. 19);

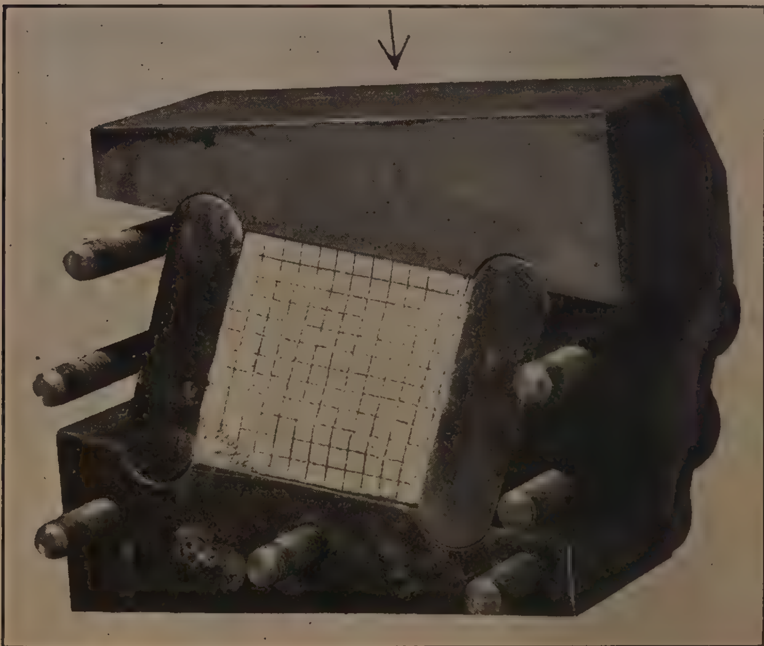

FIG. 11A. Cube of Bedford limestone being deformed by shear or rotational stress. The pressure is applied in the direction of the arrow, but is resolved along inclined shearing planes by rotation of the upper plate over the lower plate. Geologic Laboratory, University of Wisconsin.

angles between 0 and 45° are probably the more common. If compression is suspected to be the cause of a joint, it does not do, therefore, to assume that the joint developed at 45°, or thereabouts, to the pressure. It may have done so if the

strain happens to have been non-rotational, but commonly it is a mere assumption that this was the case.

In ascertaining possible relations of compression fractures to stress, it is not enough to deal with one axis of principal stress. The three principal axes of stress must be considered, or in other words, the deformation is always to be considered in three dimensions. This may be done most easily and

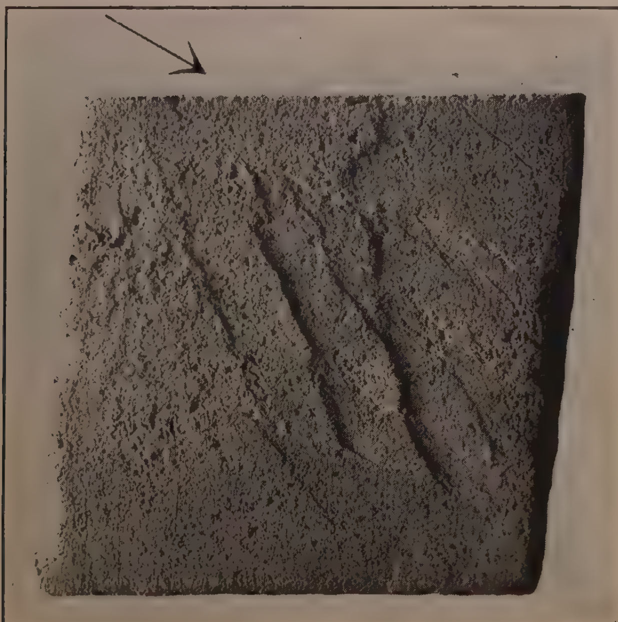

FIG. 11B. Fractures developed in Bedford limestone deformed by shearing, as in Fig. 11A. The theoretical planes of maximum shear are parallel to the sides of the blocks, but the actual fractures are at a more acute angle to the principal stress, as is common in brittle substances not heavily loaded. (See p. 35.) Geologic Laboratory, University of Wisconsin.

accurately with the use of the conventional unit of reference — the strain ellipsoid (see pp. 21–27). It will be remembered that the planes of no distortion in the strain ellipsoid are the planes of maximum shear. Any joint known or suspected to be due to compression, therefore, may be regarded as repre-

senting approximately a plane of no distortion, and the hypothetical strain ellipsoid may be imagined in relation to it. There are many possible positions in which the strain ellipsoid may be placed and still meet the requirement that a plane of no distortion be parallel to the joint in question. Remembering that the strain ellipsoid has two planes of no distortion nearly at right angles to each other, it is obvious that it can be placed so that either plane is parallel to the joint. In each of these positions, furthermore, the strain ellipsoid may be rotated through 360° on an axis normal to the plane of no distortion and still meet the requirements at any point during the rotation. Not only are there a great number of possible positions for the strain ellipsoid in relation to the joint plane, but it is to be remembered that this strain ellipsoid may be the result of either rotational or non-rotational strain or some combination of the two, — giving still wider latitude for inferences as to the actual direction of application of the stress.

Perhaps, however, the field conditions are such that the directions of shortening and elongation of the rock mass are known, and it is then possible to fix the position of the strain ellipsoid more definitely. Even then, however, it may not be possible to determine whether the strain ellipsoid is the result of non-rotational or rotational strain, and therefore whether the fracture is at about 45° to the principal pressure, or at an angle considerably less or greater. It is only in cases where the field conditions show, not only the position of all three axes of the strain ellipsoid, but in addition whether the stress was rotational or non-rotational, that the angular relations of the joint to the applied compressive stress can be stated with some precision.

In summary, joints may form under any one of the stress-strain relations (or under the conjoint action of all of them) described above as tension, non-rotational compression, and rotational compression. These are the limiting cases which cover all rock fractures. Bending and torsion, as noted later, involve merely various combinations of these relations in dif-

ferent parts of the mass.[1] In the field study of joints it is sometimes possible to determine what the stress-strain relations have been; commonly it is not. Whenever the attempt is made to relate joints to specific stress conditions, great care should be taken not to restrict the possibilities on the basis of unproved assumptions. Similarly, in naming joints, care should be taken to exclude terms which imply a more specific knowledge of the stress conditions than we possess. We see the results, not the mechanism in action.

Bending or folding, similar to the flexure of a beam or column, may involve tension on the convex side and compression on the concave side, or it may involve tension throughout or compression throughout, depending on conditions described later (p. 191). In this type of deformation joints are formed by tension and by compression (usually of the non-rotational kind), under laws already indicated.

Torsional warping. This is a kind of deformation, involving a twisting motion around an axis, which may be supposed to be common in earth deformation, but which, as a matter of fact, can seldom be identified with certainty. Folds, joints, and faults may locally be suspected to have been developed in relation to some kind of a torsional movement, but such an hypothesis is usually difficult to prove. So far as joints develop under torsional warping, they are after all subject to analysis under the limiting stress-strain relationships which have been summarized above as tension, compression without rotation, or pure shortening, and compression with rotation, or shear, just as in the case of joints related to bending or folding. Torsional warping may determine the pattern of a complex system of joints, but each joint is, nevertheless, an immediate result of tension or compression.

The conditions of strain in torsional warping are complex. They may involve either extension or contraction of the de-

[1] Likewise dilatation, positive or negative, involves expansion or compression, or some combination of them. In the Appendix references are given to experimental work on the joint pattern developed in the expansion or contraction of spheres.

formed mass, or contraction in certain parts and extension in others, with highly complex results so far as jointing is concerned. There is also in torsional warping a double cross-bending in two directions at right angles, — in each of which there may result tensional stress conditions on one side and compressional conditions on the opposite side, just as in the case of a simple beam under load in which tension is developed on the convex and compression on the concave side. In Daubree's classic experiment, in which he twisted a narrow strip of glass and obtained systematic sets of fractures at approximately 45° to the axis of torsion (see Fig. 12), the fractures obtained are evidently due to the tensile stresses involved in cross-bending. There is one set of tensional cross-bending cracks on one surface, and on the opposite surface another set of tensional cracks at right angles to the first. In a thin formation joints formed in this manner might extend through the layer, but commonly in the field, if we were to look for repetition of the Daubree experiments, we would see only one set of parallel tension cracks.[1]

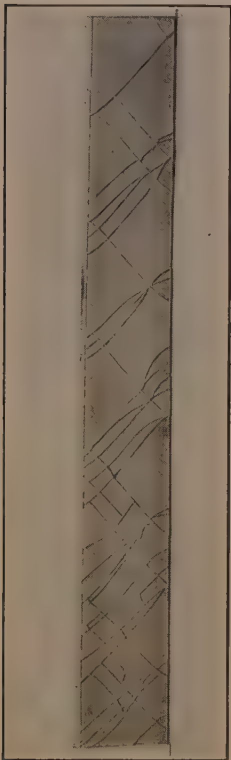

FIG. 12. Torsional warping of glass, resulting in cross-bending stresses and tension fractures. Geologic Laboratory, University of Wisconsin.

Factors modifying the theoretical relations of rock deformation to deforming stresses. In the foregoing discussion the deformation of rocks has been discussed on the assumption that they act as homogeneous bodies, and no discussion has been made of the ways in which the qualities of the rocks them-

[1] Mead, W. J., Notes on the mechanics of geologic structures: Jour. Geol., vol. 28, 1920, p. 520.

selves may tend to deflect structures from the theoretical positions assigned.

Under non-rotational compression, or pure shortening, the planes of maximum shear are theoretically at angles of 45° to the principal axis of stress. Where the substance is brittle, as most rocks are, this angle is likely to be considerably less than 45°. This has been determined experimentally. Where the material is soft or ductile the angle may be considerably more than 45°. Increase of "hoop" or containing pressures on rocks (a condition that may be expected with depth) may have the effect of making them relatively more ductile, thereby increasing the angle between the break and the principal axis of stress. In rotational strains the same variations are noted.

Rocks with structures such as bedding or igneous flow-structure or schistosity, when put under deforming stresses, are likely to fail along these preëxisting surfaces of weakness, even though these do not correspond exactly to the positions of failure required theoretically by existing stresses. As a matter of fact, however, the preëxisting structures change the positions of the surfaces of failure less than might be supposed. In the shearing of wood (Figure 9) the break occurs more or less regardless of the grain of the wood, although locally of course this has its effect.

Sets and Systems of Joints

We have considered mainly the case of a single joint. The problem of determining relations to stress for intersecting sets or systems of joints, so common in the field, is more complicated.

A group of parallel joints is commonly referred to as a set of joints. The term *set* has not been precisely defined or limited, but the common usage is as indicated. A group of two or more intersecting sets of joints constitutes a *system*. Again there is lack of precision in the term. Systems of joints may divide the rock mass symmetrically or unsymmetrically. A system may be formed as a single episode of deformation, or

as a result of successive and unrelated episodes — it is not often possible to tell which.

It is a well known fact that under experimental conditions joints tend to develop in systems rather than singly, and the geologist is constantly on the lookout to find representatives of these systems in the field. It may be said at the outset that thus far this effort has not been very successful, but that future progress in the interpretation of joints lies largely along this line of attack. The subject is a highly complicated one and we can here do no more than indicate the possible interpretations of a few simple systems.

Where two sets of joints cross at right angles it has been common practice to assume that this system represents a non-rotational strain, in which case the principal stress would bisect one of the two pairs of angles of intersection. The same result, however, might be produced in a rotational strain as may easily be seen by referring the joint planes to a strain ellipsoid and remembering that the strain ellipsoid may be produced under any combination of the limiting cases of rotational and non-rotational strain.

It has been especially common to assign intersecting sets of vertical joints to horizontal non-rotational compression. This assumption implies that the longest and shortest axes of the strain ellipsoid are horizontal and the mean axis vertical, that the direction of easiest relief and maximum extension is lateral rather than vertical. With this assumption it is not possible to place the strain ellipsoid in any other attitude. On the other hand, sets of inclined joints, intersecting each other at 90° and intersecting the horizontal at 45°, have likewise been ascribed to horizontal non-rotational compression. In this case the longest axis of the strain ellipsoid would be vertical; the direction of expansion and easiest relief would be upward. It is, therefore, apparent that horizontal non-rotational compression might produce either vertical intersecting joints or inclined intersecting joints, depending on whether the conditions allowed the easiest relief

and therefore extension of the mass, in a horizontal or vertical direction.

Many years ago Becker [1] called attention to the fact that it is mechanically possible for intersecting vertical or steeply-dipping faults and joints to develop under horizontal non-rotational compressive stresses *only* when lateral relief is easier than upward relief, and that such conditions prevail over certain areas. Many other geologists apparently have not analyzed the subject and have overlooked this qualification of Becker's, having assumed that the fact of intersection in sets implied compressive stresses, without considering alternative hypotheses, especially tension. The possibilities of lateral relief rather than upward relief are difficult to determine in the field. Irregularities of surface, like valleys, may permit easy lateral expansion in intervening ridges, thus allowing the formation of vertical intersecting faults or joints by compressive stresses; but in perhaps the greater number of cases, upward relief is easier than lateral relief.

On the assumption that vertical or steeply dipping joints and faults are the result of tension, there have been various explanations to account for their existence in intersecting sets or systems. One explanation is that they were developed simultaneously by torsion (see p. 39). Another assumes successive pulls from different directions. Still another is that they are the results of successive earthquake shocks from different directions, in which case they form under the tensional components of the waves (see pp. 271–272). Still other explanations are possible. Joints and faults formed by the cooling of an igneous mass, or the settling and drying of a sediment, may be in more or less regular sets. Relaxational settling after a period of compressive faulting or folding may develop vertical or steeply dipping joints and faults in intersecting sets. The systems in these cases are not likely to be

[1] Becker, George F., Finite homogeneous strain, flow and rupture of rocks: Bull. Geol. Soc. Am., vol. 4, 1893, p. 50.

uniform, and yet for small areas may have a considerable regularity of arrangement.

Parallel horizontal joints (with subordinate complementary vertical joints) constituting the *sheet structure* commonly observed near the surface in quarries, are supposed to be due in the main to rotational or shearing movements along horizontal planes, as indicated on page 54.

Rotational or shearing movements are supposed to account also for certain joints on the limbs of folds, one set parallel to the bedding, and the other usually curved and inclined to the bedding, in the manner described on page 185.

Horizontal shearing movement along parallel vertical planes under experimental conditions produces a complex system of joints. The joints do not develop all at once, but the different sets appear in more or less overlapping succession. Quoting from Mead:[1]

> "The first fractures to appear in any one locality on the rubber sheet are usually tension cracks inclined about 45° to the direction of the shearing movement. These are at right angles to the direction of maximum elongation and appear as vertical open cracks. They are followed immediately by two sets of vertical faults with horizontal displacement, one set striking parallel to the direction of movement and the other parallel to the free edges of the rubber sheet. These represent two directions of non-distortion or two shear planes developed by the shearing movement in which direction of relief is in the plane of the paraffin layer. Another set of faults, only two of which are shown in Fig. 14, are thrust faults, striking approximately at right angles to the tension crack and inclined approximately 45° dipping in either direction. These are due to compression in a direction at right angles to the direction of maximum elongation."

In later experiments Mead has combined shearing with nonrotational compression, probably a common condition in rock deformation. Under these conditions the first-formed tensional cracks are much less numerous and conspicuous, and both sets

[1] Mead, W. J., Notes on the mechanics of geologic structures: Jour. Geol., vol. 28, 1920, pp. 512–513.

of compressional or shearing cracks rotate in the direction of shear. In the common case of shearing along sedimentary bedding planes, there seems to be little such rotation of the cracks parallel to the bedding, probably because breaking is easier along the preëxisting planes of weakness caused by bedding than in the position exactly required by the stresses.

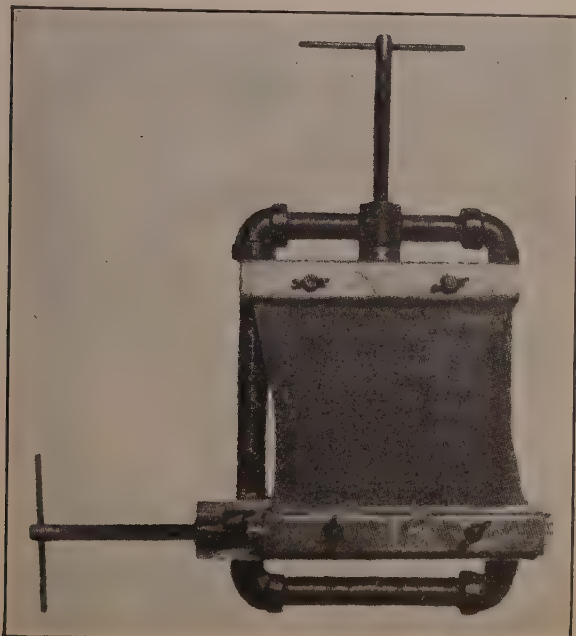

FIG. 13. Joints and faults developed by shear or rotational stress. A heavy sheet of rubber is tightly stretched between the two clamps by means of the screw at the top and coated with a thin coat of paraffin which is made brittle by chilling. The paraffin-coated rubber sheet is then deformed by means of the screw at the left. The fractures developed by the shearing movement are shown in detail in Fig. 14. After Mead.

The set of cracks inclined to the bedding, on the other hand, shows clear evidence of rotation.

Detailed study of the complex vein and fault system in the

homogeneous Butte granite suggests strongly that these structures may be due to some sort of progressive shearing movement of the kind above indicated (see p. 44).

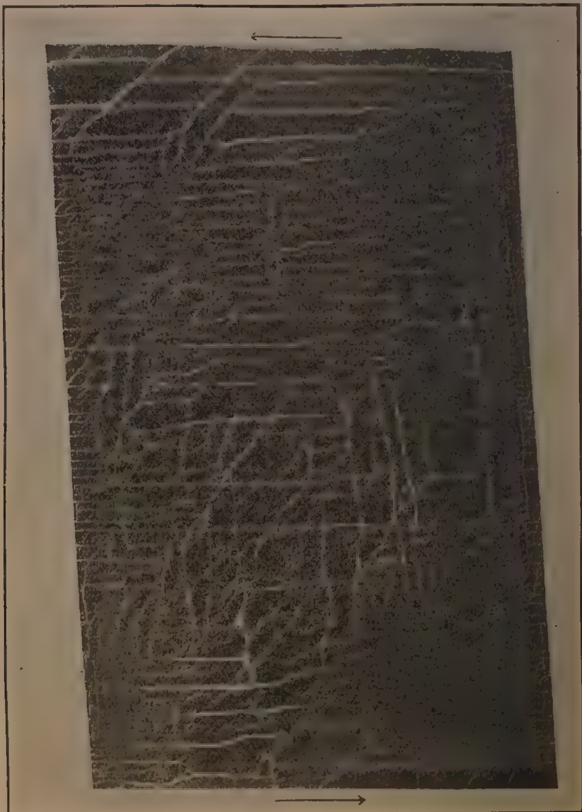

Fig. 14. Joints and faults produced in paraffin coat on rubber sheet by shearing. The arrows indicate the direction of movement and the shape of the figure shows the amount of distortion. After Mead.

In a later section on the field classification of joints, further reference is made to joint systems in relation to deforming stresses.

Apparent and Real Joint Systems

Discontinuity, vertically and horizontally, is a very characteristic feature of joints. Instead of crossing, one ends against another. The actual joint pattern is often much more irregular and complex than an idealized pattern based on the assumed extension of joints along the strike and dip. In fact, many cases have been cited of intersecting sets breaking the rock up into more or less regular polygonal forms, which have been found on close examination to be much more irregular, due to the fact that the individual joints in the field actually do not extend very far and do not often cross. There is always a temptation, in attempting to reduce natural phenomena to order, to read into a joint system a regularity which does not exist. When some suggestive system of joints, therefore, is used as a basis for inferring deformative stresses, care should be taken to examine the basis of the facts carefully to see whether the pattern actually exists or has been partly assumed.

Field Classification of Joints

It is sometimes convenient to classify joints as strike joints or dip joints, to indicate concisely their parallelism in direction with the strike or dip of beds. Again, joints may be classified as to origin, such as cooling or weathering joints. Commonly they are classified as tension and compression joints to express their relations to stresses. In nine cases out of ten, however, the student sees nothing in the joint itself which tells him whether it resulted from tension or compression, and the attempt to use this classification may lead to unwarranted conjectures, or may throw him into the discouraged state of mind of a person who believes that he should be able to tell something which the facts do not readily indicate. It is pertinent to inquire as to what conditions tell definitely whether any particular system of joints is due to tensile or to compressive stresses. These conditions are partly indicated in pre-

ceding pages, but are here presented from another point of view.

Joints which can be identified as due to tension. (*a*) Faulting may imply extension of surface (see p. 69), and hence the association of joints with such faulting would suggest their development by tensional stresses. This is not conclusive, because faulting due to tension may have followed earlier planes of weakness formed by compression.

(*b*) Open joints indicate tension, but it is difficult to determine whether tension existed at the time the joints were formed or was subsequent to their genesis.

(*c*) Tension joints have been found along the crests of anticlines, developed as indicated in the diagram (Fig. 15).

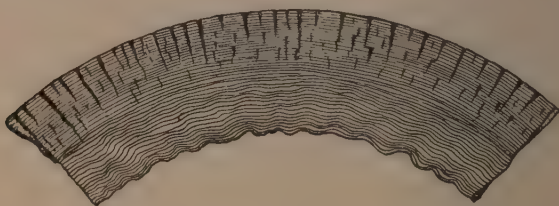

Fig. 15. Tension joints on anticline. After Van Hise.

These, however, are usually on a small scale. The writer knows of no case described for the United States in which any regional set of joints has been positively related to tensional stresses developed along major anticlines, but the existence of such a relation is plausibly inferred where joints follow the axial planes of anticlines (see also pp. 191–192).

(*d*) During the process of cooling in igneous rocks, tensional stresses are set up in them; and these stresses result in the formation of joints, not only in the igneous masses themselves, but in the adjacent rocks. The remarkably complicated fractures of Tonopah and other mineral-bearing districts of the Great Basin first suggested this origin, and it seems now to be an established fact that much, though not all, of the complex fracturing of igneous rocks may be related

definitely to their cooling [1] (see p. 72). Such joints may not be persistent nor in regular systems. Locally the fractures take certain curved or concentric forms about loci of cooling, as, for instance, in the gabbro of the Cobalt district of

FIG. 16. Jointing in Huronian slate next to contact with gabbro.

Ontario, or in the slates with which the gabbro has come into contact (Fig 16). These slates have been heated and caused to expand under the influence of the intrusive and have subsequently cracked on loss of heat. *Basaltic parting* is

[1] Sosman, R. B., Types of prismatic structures in igneous rocks: Jour. Geol., vol. 24, 1916, p. 215.

only a special type of tension jointing developed by cooling. Horizontal joints which break the vertical prisms are probably the result of shearing caused by the differential horizontal contraction of different parts of the columns during their cooling. Radial and peripheral fractures seem in some cases to have been developed by the cooling of laccoliths and batholiths. Laccoliths have sometimes been supposed to pull away from the walls in the manner of a cooling melt from a mold, as, for instance, in the Iron Springs district of Utah.

Fig. 17. Columnar basalt, Giants' Causeway, County Antrim, Ireland.

(*e*) Another type of local tension jointing is developed by the drying out of a sediment, resulting in the formation of mud cracks and of shrinkage cracks on a large scale. The joints so formed lack regularity and persistence, vertically and horizontally. It is probable that many joints in flat-lying sedimentary beds, like the Paleozoic of the Mississippi Valley, may be due to the drying and settling of the formations. The topography of the basement controls the settling to some ex-

tent, and therefore the distribution of the resulting joints. Accompanying the vertical joints caused by tension, there are horizontal joints caused by shearing, due to the differential expansion and contraction of the different layers. Mud cracks are often associated with cross breaks causing the upper layer to separate and curve up from the lower layers.

Very fine-grained homogeneous sediments like loess sometimes show considerable regularity of structure. Columnar structure similar to basaltic parting is one of the results (see Fig. 18).

FIG. 18. Cliff of loess showing characteristic columnar jointing and indistinct horizontal banding. Railroad cut near Beverly, Mo. After Hinds and Greene.

(*f*) In the very common case where rocks near the erosion surface are broken into more or less horizontal sheets by expansion under weathering, a situation in which rotational stress or shearing is supposed to be important, the subsequent settling and adjustment of the resulting thin layers of rock under gravity develop tension joints. These often follow preëxisting compression joints which were developed simultaneously with the horizontal joints. In areas of overthrust faulting and folding there is frequently evidence of

later joints and faults of a tensional nature, indicating settling or relaxational movement under gravity following the compression.

(*g*) Irregular gash cracks and veins arranged in series often show clearly an extension of the rock mass normal to the cracks. In some cases where dominant joints can be identified as the result of shearing stresses, as, for instance, in a shaly layer sheared between two hard quartzite beds, small tension gash joints have been an incidental development (see Fig. 57).

When rocks and other hard substances are broken by tension in the laboratory, or when a cement road or sidewalk cracks by tension, the break is likely to be along curved or even jagged surfaces. It is difficult to secure clean, plane surfaces of fracture extending for considerable distances; yet in the field most joints which might be ascribed to tension show these clean breaks along plane surfaces. Even in heterogeneous material such as conglomerate it is not unusual to see a clean, plane, open joint cutting indiscriminately across pebbles and matrix. If pulled apart by a tension one would expect to find projecting surfaces of pebbles. Unless there is some factor in the situation which we do not yet understand, one cannot but suspect that in these cases the joint or plane of weakness was formed by compressive shear, and that tension has merely opened up this structure. Yet it does not do to assume that all joints with clean, plane surfaces are necessarily formed originally by compressive shear, and that tension must be assigned only to jagged, irregular openings; it is known, for instance, that the formation of joints in cooling igneous rocks, as in the case of basaltic parting, develops, at least locally, smooth, plane surfaces. Fine-grained, homogeneous textures may be supposed to be more susceptible to clean breaks along plane surfaces than coarse, heterogeneous textures.

Joints which can be identified as due to compression. (*a*) Compressive joints may sometimes be identified by evidences of slipping, such as slickensides, developed along the joint

planes; but these evidences do not necessarily indicate that the compressive stresses were applied at the time the joints were formed.

(*b*) Where joints pass into overthrust faults or folds, as, for instance, in the Southern Appalachians, they are probably compression joints.

(*c*) Compression joints may develop on the concave side of a fold complementary to the tension joints on the convex side, in cases where there is a neutral layer separating zones of tension and compression. If the rock is very brittle and has low tensile strength, tension cracks may extend clear through the bed. Where the rock has considerable tensile strength, or where it is heavily loaded, compressive fractures may extend clear through the bed. While this may be theoretically a common relationship, it is seldom in the field that this condition can be identified in its idealized type form. This depends on the position of the neutral plane, discussed on pages 191–193.

(*d*) Compressive joints may also be identified frequently on the limbs of folds by the manner in which they follow closely the theoretical directions required by the conditions of compressive shear which existed at those places (see p. 185). For instance, in the Baraboo quartzite in Wisconsin (see Figs. 57 and 59), there are joints parallel to the bedding, along which there has been a slight amount of slipping; there is another set inclined to the bedding; this latter set is continuous in direction only through homogeneous beds and passes to other beds by offsetting or by curving along the bedding planes. In the softer beds the joints are so closely spaced as to yield a fracture cleavage. The joints have positions accordant with the supposition that they have been formed by compressional shearing, caused by slipping between the beds. Short open gashes or joints are also developed here by tension, as indicated in Figure 57.

A good illustration of this structure is furnished by the S-shaped joints and faults in the inclined series of quartzite

and limestone beds in the Bingham copper district of Utah, described on page 89.

(*e*) The sheet structure so commonly observed near the weathered surface in granite and sandstone is a system of jointing probably at least in part developed under compressive stresses (Figs. 19, 20 and 21). The sheets are thinnest

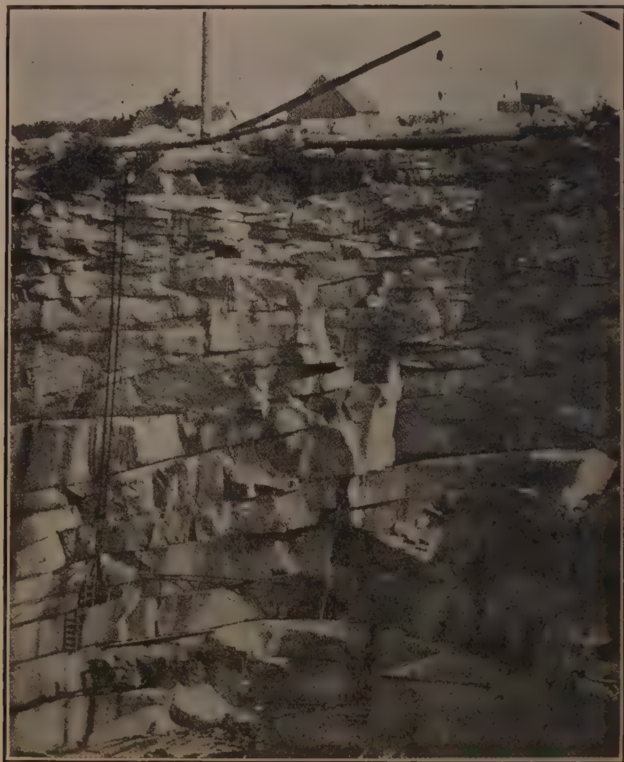

FIG. 19. Sheet structure in granite. After Dale.

near the surface and rapidly thicken below. They may be curved, and tend to parallel the rock surface whether this be horizontal or inclined. Usually they are found to be lens-shaped when traced some distance. Many instances have been noted of a lengthening of blocks when quarried out,

sometimes with explosive violence, indicating that in the ledge they were under compressive stress. Compression is indicated also by the occasional flattening, by faulting, of drill holes and other openings.[1] These compressive stresses have been referred to various causes — solar heat, weathering (or kaolinization), expansion of the surface due to removal of overlying load by erosion, and to major earth movements.[2]

The sheet structure is developed artificially by the use of explosives, by hot air, and by heating the surface. Thus when a fire is built on a ledge of granite, expansion of the surface layer often causes it to break away from the underlying rock with explosive violence; or the break may not come until the expanded surface layer cools and contracts. If a pail of cool water is thrown onto the hot and expanded surface layer it will contract so suddenly that it will part from the underlying rock. In either case the horizontal plane of parting is one of shear. At the same time the sheets are usually broken by vertical joints, which may follow planes of shear complementary to the horizontal ones, or which may be tension or bending joints formed by the settling of the flat sheets.

The sheet structure has been referred also to tension due to cooling of the igneous rocks while still under sedimentary load, the sheets being approximately parallel to the original contact surfaces of the intrusives. Bearing in mind, however, the parallelism of the sheets to the present erosion surface, and their diminution in number below the surface, the explanation of tension by cooling involves the assumption that the present erosion surface is nearly the same as the original contact surfaces, which certainly is not always true.

In beds of flat-lying sediments, like sandstone, the horizontal parting near the surface may simulate bedding very closely;

[1] Dale, T. Nelson, The granites of Vermont: Bull. 404, U. S. Geol. Survey, 1909, pp. 17–18.

[2] *Idem;* Also Dale, T. Nelson, The chief commercial granites of Massachusetts, New Hampshire, and Rhode Island: Bull. 354, U. S. Geol. Survey, 1908; and Dale, T. Nelson, and Gregory, H. F., The granites of Connecticut: Bull. 484, U. S. Geol. Survey, 1911.

in fact, many of the joints follow original planes of bedding. Close examination, however, often shows that these fractures cross the bedding at a low angle, and when followed toward a dipping erosion slope the angle usually increases. Many observations on the strike and dip of sedimentary beds taken from the natural exposure have been vitiated by the failure to recognize horizontal jointing as distinct from original bedding.

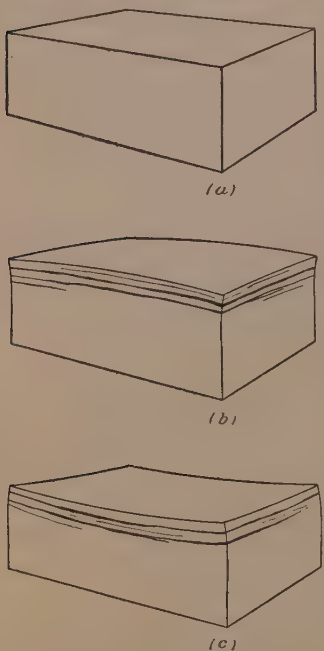

Fig. 20. Spalling of surface by shearing due to heating or cooling. After Van Hise. (*a*) Shows the condition of a block of uniform temperature. (*b*) Illustrates the manner in which the upper portion of a rock surface expands when heated above average temperature; where the difference in temperature is sufficiently great, this results in the splitting off of the upper layers. (*c*) Illustrates the contraction of the upper surface by cooling below the average temperature; where the difference in temperature is sufficiently great, this results in the splitting off of the upper layers.

The sheets are crossed by vertical joints which partly result from tension due to gravity acting on the thin sheets. Some of the joints also may be compressive. By application of the principles of breaking under rotational or shearing strain given above, it will appear that a complementary set of compression fractures should be expected approximately at right angles to the sheeting planes. In quarries these vertical joints may be in one or more intersecting sets. They are characteristically intermittent, extending through a given set of sheets and offsetting in the sheets above and below. Not infrequently they are curved. When seen in surface exposures these appear as ordinary vertical joints striking in one or more directions, depending upon whether the lateral relief was mainly in one

direction or in more than one direction. In view of the fact that processes of weathering and expansion are almost everywhere present in the surface rock, producing horizontal shears, it follows that a great many of the vertical joints we see at the surface are really related to these shears along horizontal planes and not to shears along vertical planes. In fact, the writer is inclined to assign this as one of the most common causes of vertical joints at the surface.

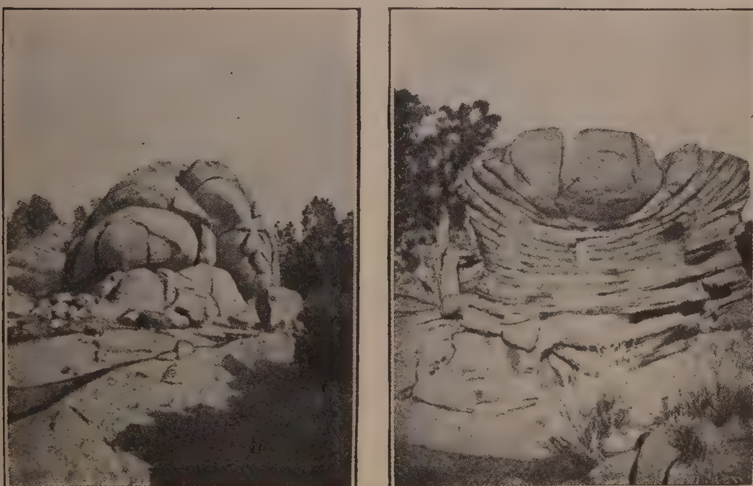

FIG. 21. Spalling of andesite outcrops, presumably due to alternate heating and cooling, in weathering.

(*f*) Closely related to the sheet structure are various curved and concentric structures characteristic of surface weathering, particularly in homogeneous igneous rocks such as granite (see Fig. 21). The rock tends to spall off at the surface in concentric shells, leaving curved and dome-shaped surfaces in the solid rock. The process affects not only the erosion surface but the surfaces of preëxisting joints, resulting in boulder-like forms of weathering. Joints so formed seem to be the result of shearing stresses acting in the same manner as in the sheet structure described in the previous section.

(*g*) Horizontal shearing parallel to vertical planes may also produce compression joints, usually in complex systems in the manner described on pages 36–37.

Depth of Joints

In general, joints are most abundant near the surface and rapidly lessen in number below the surface. This change may be sharp within a few feet; in other cases the change may extend over a few tens or hundreds of feet. The gradation is usually not uniform, especially in heterogeneous rocks. Joints may be concentrated in certain formations especially susceptible to jointing, at any distance from the surface, and the joints in other formations both above and below may be relatively fewer. Also, joints which were formed at the surface may now be deeply buried beneath unconformable terranes which are less jointed. In the deepest excavations known, extending over a mile below the surface, joints still persist, though they are comparatively few and tight as compared with those nearer the surface.

The problem of the depth of jointing is of practical importance where the joints are filled with commercial minerals. The experience of mining has shown that the vast majority of mineral-filled veins have comparatively shallow depths, that is, less than one or two thousand feet. Exceptionally, however, mineral veins have been followed down to the lowest depths to which mining has yet extended, and no one dares set the ultimate limit. However, if we were to take the evidence from mineral veins in general, we would find it bearing strongly in favor of comparative shallowness of joints.

The question of the depth of jointing may be approached in other ways — by experimental determination of the strength of rocks and inferences as to environmental conditions deep below the surface, as discussed on later pages, Chapters XIII and XIV.

Time of Jointing

An effort to ascertain the time of jointing often tends to clarify our understanding of the conditions and stresses. Where joints are numerous at the surface and die out below, more or less regardless of kind of rock or type of deformation, it may ordinarily be assumed that they have developed as surficial structures during the present erosion cycle, and inferences may be made accordingly as to conditions of their origin. Similar procedure may be followed where joints on the upper surfaces of a subjacent formation end abruptly against an unconformably overlying series. The inference then is that the jointing is related to the old erosion surface marking the unconformity.

For instance, a certain number of joints in the Baraboo quartzite, of Wisconsin, may be seen to stop abruptly at the base of the overlying Cambrian sandstone, although the sandstone is of a character susceptible to jointing. The joints are therefore pre-Cambrian in age. Others pass indiscriminately from pre-Cambrian to Cambrian rocks. They are clearly of post-Cambrian age.

In a conformable series certain beds may be extensively jointed. Others above and below are not jointed, suggesting that the joints are due to slumping and drying of particular beds, and therefore are contemporaneous with the settling of these beds. Of course, it is conceivable that the same results might be produced by tectonic disturbances applied to the series as a whole and expressing themselves principally in the weaker beds.

One of the most satisfactory ways of determining time of jointing is to note whether or not it has developed as an incident of differential movement between beds, which in turn is usually related to the folding. The time of folding can usually be ascertained. Jointing nearly always accompanies folding, so that at least part of the joints can be definitely assigned to this period. With the age of a few of the joints

determined it is often possible then to assign other joints to the pre-folding or post-folding periods.

These are only a few of the ways in which the ages of joints may be determined. There are, of course, many joints for which no satisfactory evidence of age can be found. A structural geologist cannot afford to overlook any evidence of age. It may well keep him from drawing erroneous conclusions as to the stresses and environmental conditions producing joints.

Widening of Joints by the Linear Force of Growing Crystals

Joints are often widened by solution and by mechanical erosion. A less commonly recognized agency of widening is by the forces exerted by the crystallization of infiltrated vein minerals. It has long been known that crystals in growing exert very considerable forces. Crystals of pyrite, for instance, drive apart the laminae of slates. Experiments on the pressures exerted by growing crystals of alum and other salts have shown that they exert pressures of the same order of magnitude as the ascertained resistance which the crystals offer to crushing stresses.[1] This is qualified by the fact that where an unloaded crystal is growing in the same solution the solubility of the loaded crystal is greatly increased by the pressure acting on it, and, therefore, the degree of supersaturation sufficient for the growth of the unloaded crystal is insufficient to cause growth of the loaded crystal. The mechanism by which the lifting takes place has not been demonstrated, though it has been supposed that it is caused by the expansion which takes place when the solid separates from the volume of solution which is assumed to be always under the supporting edge of the crystal.

It is supposed that this force exerted by crystals may be a factor in widening mineral-filled fissures, like the gold-bearing quartz veins of the Mother Lode of California, some of which have a width of several hundred feet. Such widths are not

[1] Becker, G. F., and Day, Arthur L., The linear force of growing crystals: Proc. Wash. Acad. Sci., vol. 7, 1905, pp. 283–288.

observed in unfilled fissures. In fact, the unfilled fissures are generally very narrow as compared with the fissures which have been filled and cemented. According to Becker,[1] laminae of the slates bounding Mother Lode veins have locally been driven apart and contorted. He concludes that when such occurrences cannot be accounted for by faulting, the inference is almost unavoidable that the laminae have been driven apart by the force of growing crystals of quartz, the axes of which stand sensibly at right angles to the planes of the laminae. The ribbon ore, consisting of parallel laminae of slate, separated by quartz, has been regarded as due to faulting, but evidence of faulting is often lacking and it is difficult to conceive how it could separate these slate bands so evenly. Separation by the growing force of quartz crystals is an alternative explanation.

Taber [2] discusses the origin of veinlets in the Silurian and Devonian strata of central New York, and concludes:

> "The fibrous veins owe their peculiar structure to the fact that the material for growth was supplied only to the base of the growing crystals through solutions occupying closely spaced capillary or subcapillary openings in the walls, while the non-fibrous veins were deposited from solutions that entered between the walls of narrow capillary fractures and bedding planes. Because of the slow rate of circulation through such minute spaces, diffusion through the solution is probably an important factor in supplying material to the growing veins."

If minerals during crystallization exert a pressure on the walls of the vein which is of the same order of magnitude as the resistance which the minerals themselves offer to crushing, as experimental evidence seems to indicate, then this force is also of the same order of magnitude as the resistance of wall rocks, and thus it becomes possible that the widening of the filled fissures may be largely due to this cause.

[1] *Op cit.*, p. 284.

[2] Taber, Stephen, The origin of veinlets in the Silurian and Devonian strata of central New York: Jour. Geol., vol. 26, 1918, p. 73.

Recrystallization of minerals in rocks, as well as in veins, is usually accomplished by volume changes, often of considerable magnitude, giving rise to pressures which locally cause notable deformation, including joints. The cumulative effect of such forces originating throughout the mass may cause earth stresses of considerable magnitude.

Surface Expression of Joints

Erosion takes advantage of fracture planes in etching the earth's surface. Joints are widened by solution and mechanical erosion. A single joint may determine the position of a

Fig. 22. Sagging of limestone beds along joints. The disturbance does not extend far below the surface. Cook's quarry (Niagara limestone), near LaSalle, Niagara Co., N. Y. After Gilbert, U. S. Geol. Survey.

great erosion trough or valley hundreds of feet wide. Where rocks are homogeneous and the fracture planes are in well-defined systems, drainage lines may be in more or less regular patterns, especially in non-glaciated regions. Where fractures

are curved and discontinuous and not in regular systems, this may be represented in the irregularity of the erosion channels. It must be remembered that fractures are not the only structures which localize erosion channels. Differing resistance of rocks, bedding, dip of impervious layers, etc., have their influence. Hence it should not be assumed that all drainage patterns correspond to fracture systems. In some cases the assumption of relationship between joints and drainage

FIG. 23. Joints in limestone widened by solution. This shows the beginning of erosion along joints.

channels has been carried so far that drainage lines have been taken as evidence of joints without further information, and continuity and regularity of joint systems have been assumed on a basis of too little information. Attention has been called above to joints of wide distribution which characteristically lack regularity. Also, joint systems in different localities are likely to have different origins and different ages. They can hardly be expected to be uniform in direction for all areas, and so far as they are uniform between localities, this can be scarcely more than coincidence.

FIG. 24. Jointing followed by erosion, Grand Canyon. After Hillers, U. S. Geol. Survey.

CHAPTER V

FAULTS

A fault is a fracture in rocks along which there has been some displacement or dislocation of one side with respect to the other in a direction parallel with the fracture. A fault differs from a joint mainly in the extent of the displacement and in the emphasis on the displacement parallel to the surface of fracture rather than normal to it. All fractures are accompanied by some displacement — fractures would not occur were not some displacement required by the stresses. Like joints, faults occur both in hard and in unconsolidated rocks, under air and under water, at the surface and deep below the surface. The following general treatment is supplemented in Chapter IX by the discussion of the recent surficial faulting seen in landslides and fresh sediments.

The preceding discussion of joints applies so largely to faults that it will not be necessary to repeat what has been said, particularly about origin and relations to stresses. The following paragraphs will be devoted mainly to the principal feature which distinguishes faults from joints, namely, the displacement or dislocation.

Naming of Faults

Some of the conventional terms commonly used in indicating the genetic elements of faults are as follows:

Fault surface is the surface of fracture. If without notable curvature it is called a *fault plane*. While most fault surfaces have some degree of curvature and irregularity, nevertheless in general they tend to approximate planes. Geologists use the term plane more often than surface, as a conventional

basis for easy reference. From a purely geometric standpoint the term plane should almost never be used. "Surface," on the other hand, fails to convey the idea of "near plane." In the following text plane is used in the sense of "near plane."

Hade is the angle made by the fault plane with the vertical. It is the complement of dip.

Dip is the angle made by the fault plane with the horizontal.

Hanging wall is the upper wall of a fault and *foot wall* is the lower wall, when the fault plane is not vertical.

Strike is the horizontal direction of the fault plane.

By *fault displacement* is meant the relative movement of the fault segments along the fault surface, which may be plane or curved. The movement may be a linear one of translation of one side with respect to the other in any possible direction parallel to the fault surface (faults of parallel displacement), or an angular movement about an axis normal to the fault surface (rotatory, pivotal or hinge faults), or some combination of these movements. Movement normal to the fault surface results merely in a wider opening, which may be common to both faults and joints. *Slip* and *shift* are also used to describe displacements.

Slip measures the actual displacement at any given place on the fault plane; that is, the distance between two formerly adjacent points on opposite walls of the fault.

Shift means almost the same as slip, but is meant mainly to indicate the displacement of rock masses as a whole, which, because of folding and local disturbances near the fault plane, may be quantitatively different from the slip measured on the plane itself.

Throw is the net vertical displacement. It is the vertical component of the slip or shift.

Heave is the horizontal component of the slip or shift. It is sometimes called horizontal slip.

Offset is the horizontal distance between the two parts of a dislocated bed measured at right angles to the *strike* of the strata.

Fault trace, *furrow* and *rift* are terms given to the line of intersection of the fault plane with the surface.

A *fault scarp* is a cliff or declivity caused at the surface by the fault movement or by later erosion which leaves the mass on one side of the fault plane standing higher than that on the other.

A *graben* or *fault trough* is a downthrow block between two upthrow blocks. A *horst* is a block upstanding between two downthrow blocks.

The definitions given above follow on the whole those proposed by a committee of the Geological Society of America in 1912.[1] However, usage is so highly varied that it is not safe to assume any finality in these terms or definitions. It is necessary to ascertain in each case what an author means.

When it comes to the naming of the fault as a whole the difficulties of naming increase because various considerations enter other than those necessary for a geometric description of the fault, — the actual direction of the displacement, whether the fault shall be described in two dimensions or three, whether the fault has caused shortening or elongation of the rocks, whether it was due to compression or tension, if compression, whether this was rotational or non-rotational, the origin of the stresses, the angular relation of the fault in strike and dip to other structures such as bedding, whether it is parallel, transverse or oblique to these structures, and so on. In view of this great variety of considerations, it is practically impossible to find names which will clearly indicate all of the characteristics of a fault and at the same time distinguish it sharply from any other kind of fault. Many names which have been used tend to emphasize one or another characteristic, and the fault may be named quite differently by an observer trying to bring out other features. A fault may be at the

[1] Report of the committee on the nomenclature of faults, by Harry Fielding Reid, William Morris Davis, Andrew C. Lawson, and F. L. Ransome: Bull. Geol. Soc. Am., vol. 24, 1913, pp. 163–186. Preliminary edition with the title, Proposed nomenclature of faults, subject to revision, was printed May 1, 1912.

same time a normal fault, a pivotal fault, a tension fault, a dip fault, and so on. This is especially true when the names

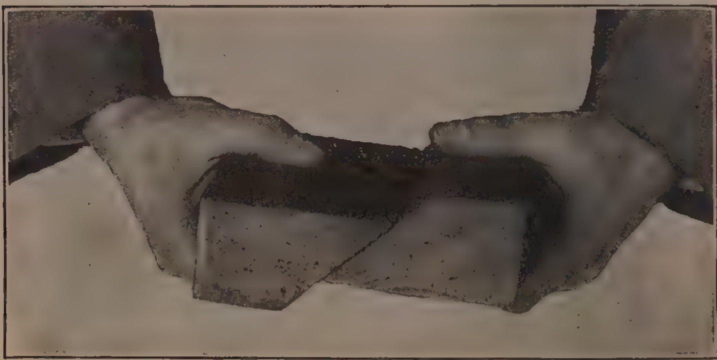

FIG. 25. To illustrate relative positions of blocks in normal or gravity faulting.

attempt to indicate origin. When we add to this the completeness of gradations between classes of faults and the complexities introduced where fault systems and not single faults

FIG. 26. Normal faulting produced by horizontal movement along table top.

are to be considered, it is clear that the accurate naming and classification of faults present an exceedingly complex problem which has not yet been solved.

Again we introduce a few fairly well standardized definitions, not as final or authoritative, but to make clear what we mean in the following pages.

A *normal* or *downthrow fault* is one in which the fault plane dips toward the downthrow side, or expressed in another way, one in which the hanging wall has been apparently depressed

FIG. 27. To illustrate relative positions of blocks in thrust or reverse faulting.

relative to the foot wall. The fault plane usually has a steep dip. This has sometimes been called a *gravity fault* on the assumption that the downthrow side has been dragged down by gravity, or *tension fault* because of the apparent extension of area. As will be seen later, the same kind of fault might be developed by horizontal movement or by thrust or compression, as well as by actual vertical depression of the hanging wall, and therefore a term implying origin or stresses may be misleading. A steep normal fault may curve below into a nearly horizontal fault, the displacement along which would be

called a thrust fault, as in the case of the Panama slides de- scribed on pages 214–216.

A *thrust* or *reverse fault* is one in which the fault plane dips toward the upthrow side; the hanging wall has apparently been elevated. Where a mass is brought forward to a notable extent by thrust faulting, more or less combined with recum-

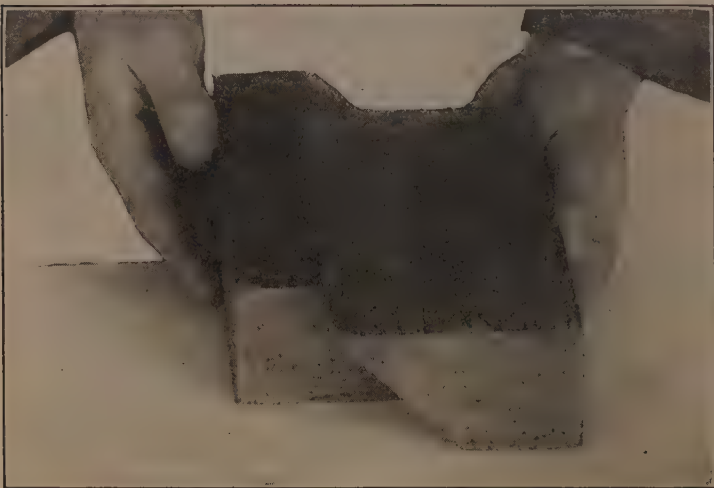

FIG. 28. Reverse or thrust faulting produced by horizontal move- ment along table top.

bent anticlinal folding, it is called by European geologists a *nappe* or *decke*.[1] (See Fig. 31.)

Where there is evidence that the hanging wall side has been pushed forward over a passive block beneath, the fault is sometimes called an *overthrust*. Where the foot wall side has been pushed under a passive hanging wall it is called an *underthrust*. Overthrusts and underthrusts produce the same kind of displacement, and it is not often possible to determine whether the result has been accomplished by overthrusting or underthrusting or both. In fact, the use of these terms

[1] Bailey, E. B., The structure of the southwest Highlands of Scotland: Quart. Jour. Geol. Soc., vol. 78, 1922, p. 87.

usually implies some hypothesis or assumption on the part of the observer.

A *heave fault* or *horizontal slip fault* is one in which there has been horizontal movement along a fault plane which is usually highly inclined.

Normal, reverse, and heave faults are all faults of *parallel displacement*, meaning that all straight lines on opposite sides

FIG. 29. To illustrate hinge faulting. This would appear as a normal or gravity fault on a plane normal to the fault plane passing through the ends of the blocks nearest the reader and as a thrust or reverse fault in a plane passing through the ends of the blocks farthest from the reader.

of the fault and outside the disturbed zone which were parallel before the displacement, are parallel afterward.

A *hinge fault* is one in which the movement is an angular or rotational one on one side of an axis normal to the fault plane. The movement is all in the same direction but increases in amount away from the axis. A *pivotal fault* is also

an angular or rotational one about an axis normal to the fault plane; the movement on opposite sides of the axis is in different directions. It is clear that either a hinge or pivotal fault, if inclined, may produce displacements which, in a given cross-section, might appear as normal or reverse faults.

Nearly all faults have some rotatory movement, but they are not classified as rotatory unless this movement is conspicuous. Rotatory faults have not received much attention in geologic literature. Some of the best examples are in areas of surface volcanics in the western United States, where they have been mapped in connection with ore deposits (Tonopah, Iron Springs, Bullfrog, etc.).

The rotatory movement here described should not be confused with rotational strain (see p. 19). Rotational strains are involved both in rotatory faults and in faults of parallel displacement.

In stratified rocks it is sometimes desirable to emphasize the relation of the fault to the bedding by calling it a *strike fault*, or *dip fault*, or *oblique fault*, or *longitudinal fault*, or *transverse fault*, or *transcurrent fault*, or *bedding fault*—terms which are self-explanatory when it is remembered that they refer to the manner with which the fault plane parallels or cuts the bedding. Faults parallel to bedding are often the most difficult to recognize, but they are coming to be known as numerous and important.

There are additional names applied to systems of faults, defined on page 85.

Some of the best known types of faulting are discussed in the following pages.

Faults Associated with Igneous Rocks

Faults are likely to be numerous within and adjacent to areas of igneous activity. They are especially numerous in surface volcanics. Such faults are usually irregular and discontinuous, and offset along cross-faults and joints, breaking the rocks into heterogeneous polygonal blocks. Displacements are both

horizontal and vertical. While normal faults predominate, all other kinds are represented. These faults are well illustrated on many maps of western mining districts prepared by the U. S. Geological Survey, notably those of the Tonopah,[1] Goldfield,[2] Bullfrog,[3] and Clifton [4] districts.

It has long been suspected that there is some genetic connection between faulting and igneous activity. Spurr expressed this specifically as follows: [5]

"It is plain that the faulting was the result of adjustments of the crust to suit violent migrations of volcanic rock; that it originated with the swelling up of the crust and its forcible thrusting up and aside to make way for the numerous columns of escaping lava; and that after the cessation of the eruptions it was continued by the irregular sinking of the crust into the unsolid depths from which the lavas had been ejected. It can readily be seen that all sorts of pressure (from below upward, lateral, and downward, by virtue of gravity) must have been concerned in such movements, and that the first faults were due rather to upward and lateral irregular thrusts, while the later ones (in many cases along the same planes as the first) were due to gravity. So reversed and normal faults are equally natural, and both occur frequently."

"The writer at first looked upon the faulting at Tonopah as exceptional and local, and not to be connected with ordinary faulting in the Great Basin, but there now appears no reason for doubting that the phenomena within this small, carefully studied area are typical of the unstudied similar volcanic region beyond the limits of the map."

In discussing joints, attention has been called to the common development of joints and partings during the cooling of igneous rocks, including peripheral, radial, concentric, basaltic, and irregular partings, — both in the igneous rock itself,

[1] Spurr, J. E., Geology of the Tonopah mining district, Nevada: Prof. Paper No. 42, U. S. Geol. Survey, 1905.
[2] Ransome, F. L., Geology and ore deposits of Goldfield, Nevada: Prof. Paper No. 66, U. S. Geol. Survey, 1909.
[3] Ransome, F. L., Emmons, W. H., and Garrey, G. H., Geology and ore deposits of the Bullfrog district, Nevada: Bull. 407, U. S. Geol. Survey, 1910.
[4] Lindgren, Waldemar, Copper deposits of the Clifton-Morenci district, Arizona: Prof. Paper No. 43, U. S. Geol. Survey, 1905.
[5] *Op. cit.*, p. 80.

and in the adjacent rocks. Faulting may follow any of thes surfaces of weakness.

Intrusive igneous rocks, particularly great plutonic masses cause faulting and other deformation in the adjacent rocks b means of pressure exerted during intrusion, by the expansio due to heating, and by the cooling which follows. Th

FIG. 30. Mosaic pattern of faults in Tertiary lavas of the Tonopah district. Note lack of continuity of faults. After Spurr. From Cleland's Geology, by arrangement with American Book Company.

common association of plutonic intrusion with regional defor mation of a mountain-making kind is one of the salient fact of structural geology (see pp. 249–252). The upthrust o igneous intrusion is in some cases supposed to cause norma step-faults in the overlying rocks, but usually the faulting i more complex.

Igneous intrusion not only causes faults but may be in par an effect of faulting; at least the planes of weakness ofte localize intrusion. Dikes or sills are often intruded along faul planes. In some of these cases there has been movement afte the intrusion, as shown by the shearing or faulting of th dikes themselves.

In general, regions of active vulcanism are likely to b regions of faulting, and *vice versa*. Linear distribution o vulcanism has even been used as evidence of faulting. Whil we have indicated certain type relationships of faults wit

igneous rocks, the full story in all its implications is yet to be worked out.

Faults Associated with Ore-Bodies

Faults have a close relation to many ore deposits. Not only may fault fissures be filled with ore, but the ore-bodies themselves may be displaced by later faults. Faults related to igneous activity have an especially close relationship to ore-deposition which has been well summarized by Spurr.[1] As a result of analysis of conditions in many mining districts where the ores are related to igneous activity, he concludes that in most of the cases the veins are formed in distinct fissures, often of great length and depth, but with very slight differential movement or faulting, and that the faulting of greater magnitude "began directly after or during ore-deposition, and developed slowly afterward, through very long periods of time, in many cases down to a geologically recent period; and the later fault-fissures are uncemented by vein or gangue material." He further indicates "that in most of the cases mentioned the faulting accompanied a local domal uplift affecting a restricted area, which roughly coincided with, and seemed to have a location connected with, the area of earlier mineralization." As a reason for this observed order he assigns the progressive and cumulative effects of cooling, which continues long after ore-deposition has ceased. The late doming he regards as due to pressure of the intrusion from below, which expresses itself in the surficial rocks only after erosion has sufficiently lessened the load and resistance to upward thrust.

Normal Faults in Unfolded Sediments

Normal faults may be locally developed in nearly flat-lying sediments. Here the cause of tension may be shrinkage and settling due to drying and recrystallization. A displaced layer

[1] Spurr, J. E., The relation of ore-deposition to faulting: Econ. Geol., vol. 11, 1916, p. 615.

is sometimes covered by a continuous layer of sediment, suggesting that the faulting was contemporaneous with deposition. Often no other causes are discernible, but it is not possible to exclude hypotheses of regional or deep-seated tension related to major earth movements.

Association of Normal Faults with Folds

A normal fault may pass into a fold along the strike, down the dip, or even up the dip. The Kaibab fault of the high plateaus of Utah is an illustration of a normal fault which grades along the strike into a monocline.

Normal faults associated with overthrust folds, to be seen for instance in the southern Appalachians, seem to be the natural consequence of settling following the disturbance of equilibrium by thrust, — in other words, of relaxation so commonly following compression.

The attempt has been made also to correlate normal faults existing over a large area with the collapse of a very gentle arch. To illustrate, the great normal faults in the Great Basin area are referred to the collapse of an arch originally extending from the Wasatch on the east to the Sierra Nevadas on the west.[1] Where broad, gentle arches are thrown up through compression or through changes in support below, the inherent weakness of the rocks may cause them to break almost from the start, and allow certain blocks to settle within the arch. Chamberlin[2] has called attention to the inherent weakness of rocks and their inability to support themselves in large masses. A dome corresponding to the sphericity of the earth, of strong crystalline rocks, of any extent and thickness, would, if unsupported below, sustain only 1/525 of its own weight. It is therefore apparent that when any great movement is initiated tending to arch any part of the earth's surface, unless this arch is thoroughly and evenly supported by great masses below, it will be unable to sustain

[1] Le Conte, Joseph, On the origin of normal faults and of the structure of the Basin region: Am. Jour. Sci., vol. 38, 1889, p. 262.

[2] Chamberlin, T. C., and Salisbury, R. D., Geology, vol. 1, 1904, p. 555.

itself by its own strength alone; and one would expect a settling of blocks, giving the tension or normal type of faulting and jointing, with consequent extension of surface.

No such relation as that discussed in the above paragraph has been proved on any large scale. The existence of such a primary arch or tendency for arching is inferred as a possibility from the existence of normal faulting.

Another possible effect of arching of a competent bed is to cause normal faults in less competent beds above, as if a plug had been driven up from below.

Normal faults, as well as other kinds of faults and joints, are supposed to be characteristic developments along the axes of moving geanticlines of the kind described by Brouwer in the Dutch East Indies.[1] The movement is not only one of bending but of horizontal migration of the axis of the fold.

Thrust or Reverse Faults

A *thrust* or *reverse fault* is one in which the fault plane dips toward the upthrow side. This kind of faulting indicates apparent shortening of the rock surface and therefore compression; but as in the case of normal faulting, the use of these terms implying origin or actual displacement may be misleading because similar faults may be formed in other ways (p. 94). Closely spaced thrust faults may slice rocks, giving an imbricate structure or distributive faulting.

Thrust faults characteristically have low dips, usually less than 45°, or even nearly horizontal. Because of the low dip and the inequalities of the erosion surface the fault trace at the surface is likely to be an extremely irregular one. In the Rocky Mountain region, because of erosion of the overlying block, the fault trace may appear even as circular.

R. T. Chamberlin discriminates low-angle overthrusts approaching horizontality from ordinary reverse faults of higher dip, though he recognizes that the two are closely associated

[1] Brouwer, H. A., Fractures and faults near the surface of moving geanticlines: Proc. Kon. Akad. van Wetenschappen te Amsterdam, vol. 23, 1920, pp. 570–576.

in origin and that they are connected by composite types. As a result of experimental work he lists factors which will tend to lower the angle of faulting.[1]

Thrust faults may develop as rotational or non-rotational strains (see pp. 18–19), but it is supposed that they are dominantly the result of rotational strains or shear. Low-angle thrust-faulting by shear requires theoretically the development of a complementary set of fractures dipping steeply toward the direction from which the overthrust comes. Nearly all thrust faults are accompanied by many fractures with a position corresponding to this requirement. Along many of these fractures there has been normal faulting. This may have taken advantage of the fractures formed during compression and occurred after the compression was over, or the faults may be direct gravity breaks during the later relaxational period when the overthrust block was readjusted and settled under gravity.

Thrust faults are often associated with recumbent or drag folds, as in the Alps, the Scottish Highlands and the southern Appalachians. The mass brought forward by the combination of faulting and folding is called a *nappe*. The fault planes may be inferred to have the relations to stress indicated on page 19 in rotational or shearing compressive strains. The inference is usually made that there is an overthrust, and therefore shortening, of so many feet. This is true in the plane of the cross-section; it tells us nothing of the movements inclined to the plane of the section, which may have been fully as great. Consideration of many cross-sections is the same in effect as considering the fault in three dimensions, and leads to closer estimates of actual shortening.

[1] Chamberlin, R. T., and Miller, W. Z., Low-angle faulting: Jour. Geol., vol. 26, 1918, pp. 1–44.

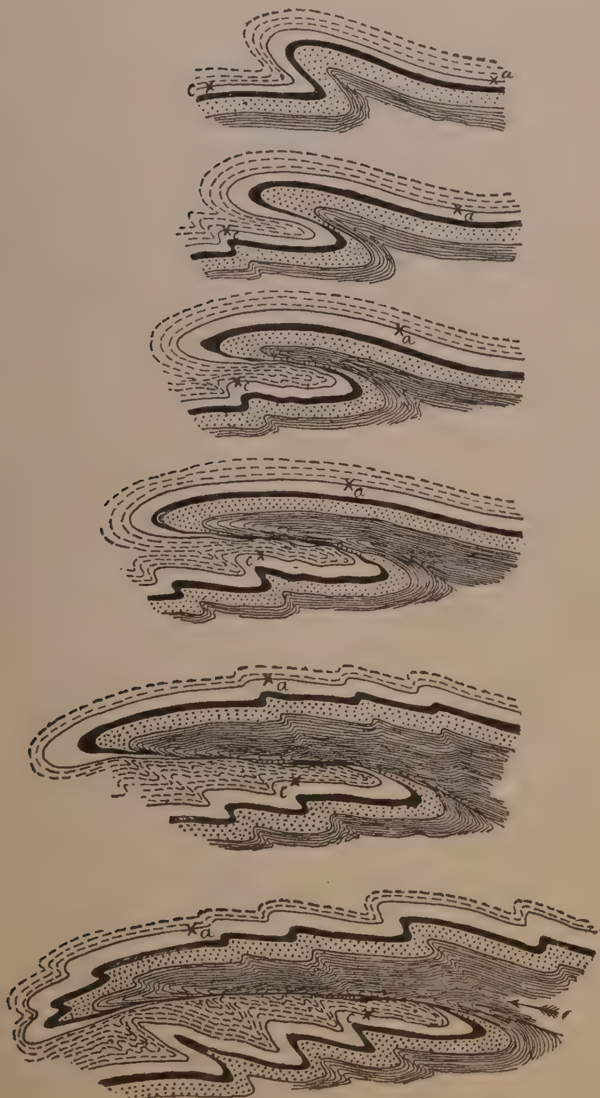

FIG. 31. To illustrate development of overthrust folding and faulting (nappe), accompanied by minor drag folds, as inferred from Alpine structure. After Heim.

An examination of the United States Geological Survey folios brings out the interesting fact that in the southern Appalachians 83% of the thrust faults, as indicated on the

FIG. 32. Idealized sketch by Heim showing hypothetical relations of thrust faults in the northwest Highlands of Scotland to the overthrust folds and faults of the Alps. The sketch is intended to bring out the idea that the northwest Highlands structure represents the roots of mountains which have been removed by erosion.

cross-sections, are definitely related to overthrust folds. Willis[1] classifies them as (1) *break thrusts*, where the thrust-

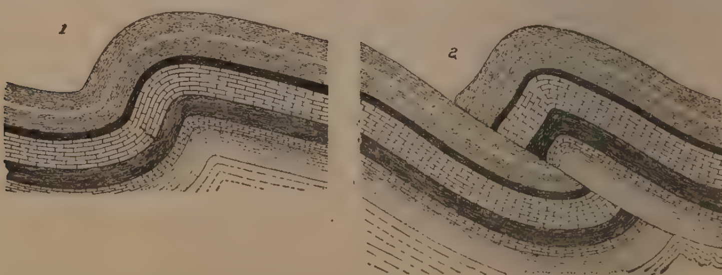

FIG. 33. Overthrust faulting localized by tension fracture; "break thrust." After Willis. 1. Shows break in the massive limestone bed which determines the plane of the break thrust along which the displacement shown in 2 takes place.

fault plane follows a previously formed tension fracture on the crest of the anticline; (2) *shear* or *stretch thrusts*, where the

[1] Willis, Bailey, Mechanics of Appalachian structure: 13th Ann. Rept., U. S. Geol. Survey, pt. 2, 1893, pp. 222–223.

break follows the sheared and stretched underlimb of an overturned fold; and (3) *erosion thrusts*, where the competent layer carrying the thrust is first weakened by erosion at or near the crest of the anticline. (Figs. 33 and 34.)

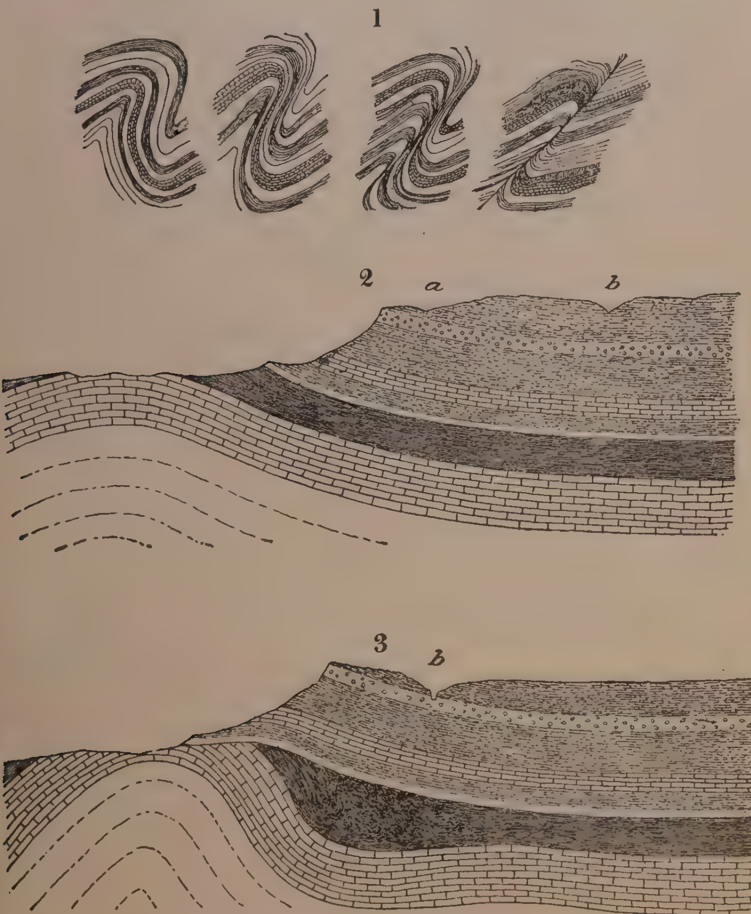

FIG. 34. Illustrating thrust fault developed by stretching and by erosion. After Willis. 1. Stretch thrust developed from an overturned fold by stretching of the middle limb; 2. Erosion profile and section of a simple anticline; 3. Erosion thrust developed from the condition shown in 2 by compression from the plateau side, accompanied by continued erosion.

Geological literature contains a rapidly increasing list of great thrust faults bounding the bases of mountains and dipping into them from one or both sides. Especially notable has been the recognition in recent years of overthrust faults in the Rocky Mountain region of Utah, Wyoming, and Montana. It is possible that with the progress of field work many of these faults will be correlated with one or a few great faults, but it is also known that in some places the fault movement is not along a single plane but is distributed among several. The wide prevalence of faults of this sort related to mountain ranges suggests that this kind of faulting is an important factor in mountain-making, just as normal faults have been so regarded in relation to the great Basin and other ranges (see pp. 103–104).

The obviously greater displacement along the low-angle and horizontal thrust planes as compared with the high-angle thrust planes has been a matter of comment. There are few vertical displacements which reach a mile, whereas the horizontal displacements are measured sometimes in tens of miles. This fact alone requires emphasis on tangential shearing parallel to the earth's surface as a cause, and leads naturally to the hypothesis that the surface rocks have moved horizontally over a rigid mass below.

It has ordinarily been assumed that great thrust faults are developed at considerable depths and that they are exposed now because they are uncovered by erosion. Goldschmidt,[1] however, in studying the great Caledonian thrusts in Norway, finds evidence that conglomerates have formed contemporaneously with the thrusting in advance of the scarp of the overthrust mass, and have in turn been overridden and sheared by the fault. If this is true, the fault must actually have emerged at the surface. There is no reason why a thrust fault should not emerge at the surface, as well as a normal

[1] Goldschmidt, Victor M., Om hoifjeldskvartsen I og II: Norsk Geologisk Tidsskrift, vol. 4, pt. 1, 1916, pp. 44–46, 49–53. Reviewed in Geol. Mag., vol. 4, 1917, pp. 130–132.

fault, but the actual evidence of this emergence during the faulting (not by later erosion) has been extremely limited.

A classic case of renewed faulting marked by disturbance of conglomerates is the great Keweenawan fault of Michigan described by Irving and Chamberlin.[1]

So far as the relationship of the overlying mass to the underlying mass along a thrust plane is concerned, the fault might just as well be called an underthrust fault and the fold associated with it might be called an underdrag or underthrust fold. The term overthrust involves the conception that the overhanging mass has moved forward over a passive mass below, and the use of the term underthrust implies that the underhung mass has been the active and moving one. Some geologists have been inclined to think of underthrusting as an important orogenic process. It is seldom possible to be sure whether a so-called overthrust fault may not really be an underthrust, but probably most geologists incline to the view that most of these occurrences really represent overthrust, — on the ground that surficial rocks are known to be more easily deformed and to move more easily than underlying rocks.

Thrust faults with miles of displacement may cause but little local disturbance in the rocks along the fault plane, and little gouge or brecciation. It is hard to understand how there can be so little disturbance. In other cases there may be considerable gouge and extensive brecciation extending for many tens of feet into the rock above or below, and folding, warping, and jointing may be common. In still other cases a secondary schistosity or cleavage is developed nearly parallel to the fault plane as a result of drag, and this schistosity is almost always related to drag folds, the axial planes of which are parallel to the schistosity. The increase in schistosity toward the fault plane is clear evidence of its contemporaneous development, though in some cases preëxisting schistosity may

[1] Irving, R. D., and Chamberlin, T. C., Observations on the junction between the Eastern sandstone and the Keweenaw series on Keweenaw Point, Lake Superior: Bull. 23, U. S. Geol. Survey, 1885.

have afforded the plane of weakness of which the fault movement has taken advantage.

Thrust faults involve fracture alone, or fracture combined with rock flowage. It is notable that where rock flowage accompanies the thrust-fault plane the fault is likely to be of a minutely distributive type (see Fig. 35). There is no sharp line of demarcation between a structure which in general displacement effects might be called a thrust fault, and a cleavage structure which might be called a slip zone of rock flowage.

How far distinct faults can go below the surface before they merge into a zone of rock flowage, or whether there is any lower limit, is a problem like that for joints, already referred to (p. 58), and is further alluded to in connection with a subsequent section on curved fault planes.

Heave or Horizontal Slip Faults

Under this heading are considered horizontal displacements along vertical or nearly vertical fault planes. If the beds are horizontal, such horizontal movement will cause neither upthrow nor downthrow; if they are inclined or folded, however, such movement may cause stratigraphic downthrow on one part of the fault plane and upthrow on another, except when the fault is parallel to the strike.

The faulting in which the California earthquake originated followed a vertical plane at least 600 miles long. Along this plane the maximum horizontal movement was about 21 feet, while the vertical movement was only a small fraction of this. This is one of the few cases of definitely proved horizontal displacement.[1] (See p. 105.) Illustrations on a much smaller scale may be found in the faulting of the igneous rocks of many western mining districts, cited on pages 72–74. Striations on fault surfaces not uncommonly show that there has been a component of horizontal displacement, even though

[1] Gilbert, G. K., The earthquake as a natural phenomenon: Bull. 324, U. S. Geol. Survey, 1907, p. 4.

the major displacement is vertical. More attention is now given than formerly to possibilities of horizontal displacement, with the result that more information in regard to this type of movement is becoming available. As yet, however, good illustrations are few.

Major and Minor Faults and Fault Systems

Thus far we have dealt with faults mainly as separate entities. It remains to consider some of the problems involved when faults are grouped, as they commonly are, into systems, known as *multiple faults, major and minor faults, major and auxiliary faults, shear zones, distributive faults,* etc. It is often much more essential to understand the nature of the system than to classify and work out the actual displacement and origin of an individual fault.

The determination of the actual displacement of a fault may yet fail to disclose the major faulting or major deformation of an area. Normal faults with vertical displacements may be incidental accompaniments of a great thrust fault. In almost any complexly faulted area the local displacements may be varied and yet the major and controlling displacement may be a comparatively simple one. For local purposes, for instance in working out the structure of an orebody, the determination of the actual displacement of a single fault may suffice; but usually it is desirable to know the relation of these local displacements to any major structure. With the larger features of movement in view it becomes easier to interpret and correlate a variety of local faults. Thus it is that fault systems or patterns require much attention from the structural geologist.

What has been said about systems of joints on a preceding page applies equally well to faults, but additional mention may be made of a few common fault systems.

It is seldom that a great thrust fault is unaccompanied by normal faults displacing the thrust-fault plane. The normal faults may run in any direction with relation to the strike

of the thrust fault. This association has been explained on the ground that the thrust faulting, under compression, has been followed by a relaxational period during which gravity causes a settling of the masses, in order to regain the equilibrium disturbed by the compressive horizontal stresses. The initial cracking along nearly vertical surfaces may have taken place during the thrust faulting under compressive stress and have opened up later, or the cracking itself may have taken place later. Theoretical considerations favor the complementary development of simultaneous cracks of this kind.

A thrust fault may be one of a series of more or less parallel faults which may slice an area into thin layers. This is *imbricate structure* or *distributive faulting*. In such a system movement along one plane may greatly predominate over movement along other planes, in which case the other faults may be regarded as subsidiary expressions of a single great movement. Systems of this sort may be closely associated with overthrust folds, as they are in the southern Appalachians, or they may be nearly free from folds, as in the Highlands of Scotland.

In the northwest Highlands of Scotland the beds are minutely sliced and piled one on top of the other. As the

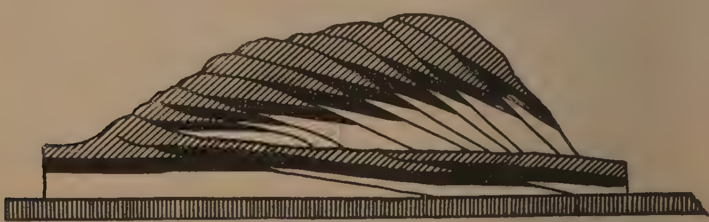

FIG. 35. Major fault plane or fault sole. After Cadell.

deformation continues these beds may ride forward as a group over a major fault plane at the bottom, sometimes called the "sole." The reports and maps of the British Geological

Survey [1] on the Scottish Highlands afford an unrivaled opportunity for the study of faults of this type. Experimental reproductions of these faults by Cadell [2] throw light on the process (Fig. 35). Some of his conclusions are quoted:

1. Horizontal pressure applied at one point is not propagated far forward into a mass of strata.
2. The compressed mass tends to find relief along a series of gently-inclined thrust-planes, which dip towards the side from which pressure is exerted.
3. After a certain amount of heaping-up along a series of minor thrust-planes, the heaped-up mass tends to rise and ride forward bodily along major thrust-planes.
4. Thrust-planes and reversed faults are not necessarily developed from split overfolds, but often originate at once on application of horizontal pressure.
5. A thrust-plane below may pass into an anticline above, and never reach the surface.
6. A major thrust-plane above may, and probably always does, originate in a fold below.
7. A thrust-plane may branch into smaller thrust-planes, or pass into an overfold along the strike.
8. The front portion of a mass of rock being pushed along a thrust-plane tends to bow forward and roll under the back portion.
9. The more rigid the rock, the better will the phenomena of thrusting be exhibited.
10. Fan-structure may be produced by the continued compression of a single anticline.
11. Thrust-planes have a strong tendency to originate at the sides of the fan.

In the Butte district of Montana the copper-bearing veins occupy faults of three different ages. The first and most heavily mineralized is an east-west set, the second, less mineralized, is a later northwest set, and a third, only slightly

[1] Peach, B. N., Horne, John, Gunn, W., Clough, C. T., and Hinxman, L. W., The geological structure of the northwest Highlands of Scotland, with petrological chapters and notes by J. J. H. Teall, edited by Sir Archibald Geikie: Mem. Geol. Survey of Great Britain, 1907.

[2] *Op. cit.*, pp. 473–476.

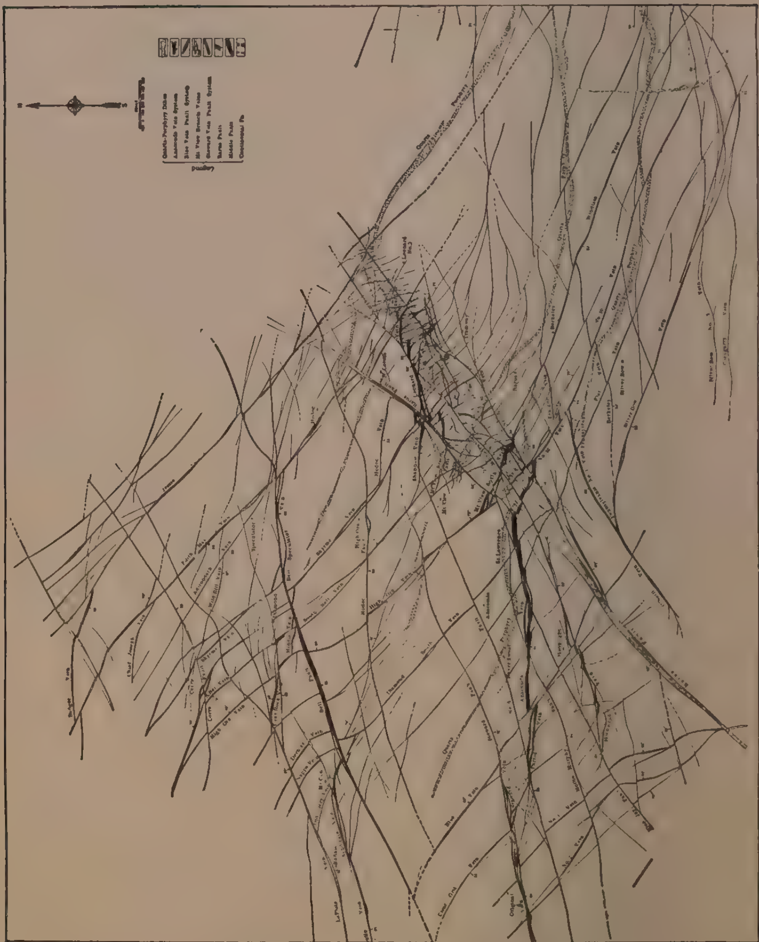

FIG. 35A. Horizontal plan of fault and vein systems at Butte, Montana, showing successive and overlapping development of the east-west, northwest and northeast faults and fissures. After Sales.

mineralized, is a northeast set. The later ones intersect the earlier ones. The exploration and development of this camp, as well as litigation concerning apex rights, have required the most intensive geologic study of these faults. In the past

these sets of faults have been treated as separate structural units developing successively. It now appears that the mineralization accompanying them is substantially of one period, indicating that the successive structural movements took place mainly within the period of mineralization. It further appears that the clean cut separation into structural units fails to take into account the existence of many branches, cross-overs, and gradational features tending to tie the different systems together. The first two are probably somewhat complementary, and related to a single major and progressive shear of the district of the kind illustrated in Figure 35A.

In the Bingham district of Utah a gently folded quartzite series, with thin layers of limestone, is cut by many faults, some of which are mineralized by copper, lead and zinc ores, and some of which are later than the mineralization. The mineralized faults are parallel to the strike or dip of the bedding or in intermediate directions, and at first might seem to be separate and unrelated units, but a correlation of all of the faults brings out certain illuminating features. Many of them are curved, are co-terminal with certain beds, pass into the beds tangentially, and in general have the relation to bedding indicated in Figures 57, 58, and 59 which is characteristic of fractures developed by shear accompanying differential movement between the beds. Others of the faults may have a different origin. The consequence of the recognition of system among at least an important part of mineralized faults is the conclusion that the extension of any fault cannot safely be projected along a plane surface or from bed to bed, and that faults in different beds cannot be safely correlated across an unexplored area. The bearing of this on the problem of determining the apex of a given mineralized fault, found far below the surface, is obvious.

Correlation of Faults

Unless the actual displacement and the general habit of the fault system are known, it is hardly legitimate to infer ex-

tensions or correlation of individual faults between separated areas. For some major faults this is possible, but in complexly faulted regions the fault directions are seldom sufficiently uniform to warrant their correlation with faults of substantially the same directions in other districts. Especially is the extension and correlation of faults unwarranted in regions of igneous rocks, where, as shown by the various maps of western mining districts (such as those of the Tonopah, Clifton, Globe and Bisbee) faults run in nearly all directions, intersect at all angles, change their directions, are cut off suddenly, and in fact, show all the irregularities to be expected from interior strains of intrusion and cooling. One is scarcely warranted in any of these camps in extending a fault on the map more than a few feet beyond where definite evidence of it is seen, for it may suddenly end or change its direction entirely. Scarcely less irregular are the faults caused by drying and settling of sediments. When one considers the heterogeneity of rocks taken on a large scale, it is to be expected that, even though stresses be applied in a uniform direction over a large area, these stresses will be transmitted and resolved in such directions and intensities as to develop fractures with great variety of attitudes. Hence the difficulty of correlating faults over wide areas or between heterogeneous systems of rocks can scarcely be overestimated.

Moreover, after rocks have been fractured they may be deformed by folding, in which case the fault and joint planes may be so distorted that they will appear on the surface as curved lines. The folded thrust-fault planes in the southern Appalachians illustrate the remarkable complexity which may be developed in a joint or fault plane. Topographic irregularities cause a fault plane with low dip to appear curved on the surface. The surface distribution of such folded faults has little similarity to the idealized sets of straight-line intersecting faults often presented as typical of fault conditions.

Curved and Folded Faults

Curved fault and joint surfaces, especially joint surfaces, may be formed by spalling of rocks due to insolation, and by other processes. Fractures related to the cooling of igneous rocks may be curved. Initially curved fractures are found in other relations where it is not easy to analyze causes. They may be due to the nature and application of the stresses, or to lack of homogeneity of the rock.

Many great low-angle thrust-faults seem to have an initial curvature, although it is often not easy to discriminate this from the results of later folding.

In some cases the thrust-fault surfaces steepen toward the erosion surface. Experimental work on low-angle faulting under certain conditions likewise shows such a steepening near the surface.[1]

The idea that thrust-fault planes commonly flatten below finds favor because of a more or less general consideration, namely, that the earth is known to be highly rigid below the surface, and less rigid near the surface; that deformation of rocks is largely superficial, and that tangential movements in the earth's surface involve a horizontal slipping or shearing of a frangible shell over a more or less immovable core, which would form horizontal faults, often simultaneously with and nearly parallel to a cleavage caused by rock flowage. In harmony with the idea is the fact that displacements along nearly horizontal fault-thrusts are much greater than along fault planes of steeper angle, ranging up to ten miles and in some cases several times this figure.

While many field observations are in accord with the hypothesis of the flattening of thrust faults with depth, they are nevertheless far from sufficient to constitute a sound basis for inductive generalizations. Very few fault planes are sufficiently exposed to determine whether they do flatten with

[1] Chamberlin, R. T., and Miller, W. Z., Low-angle faulting: Jour. Geol., vol. 26, 1918, pp. 1–44.

Quirke, T. T., Concerning the process of thrust faulting: Jour. Geol., vol. 28, 1920, p. 417.

depth; and even observed cases of flattening may be merely local warping, and still deeper erosion might show steepening. At present no positive conclusion can be reached. If a cube of soft clay, held stationary at the ends and with room for escape upward, be compressed by non-rotational stress from one side, a thrust fault will be developed on its upper side dipping toward the thrust. Lower in the cube this thrust fault steepens and grades into a plastic deformation which approximates rock flowage.

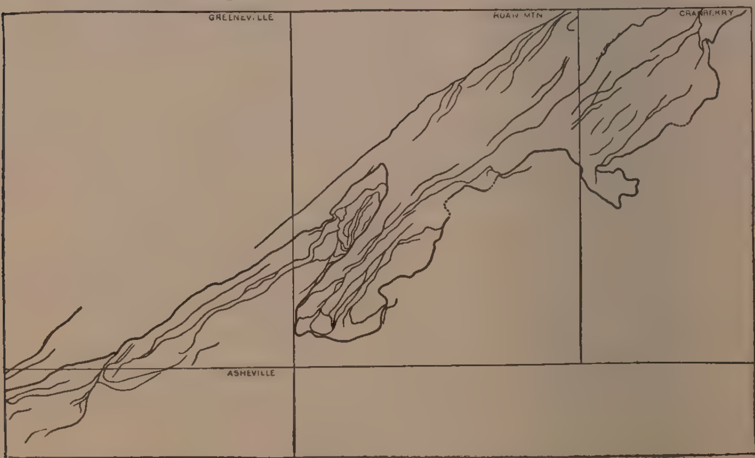

FIG. 36. Map of the faults in the Roan Mountain and adjacent quadrangles, Tennessee and North Carolina, showing the relation of the minor faults (lighter lines) to the earlier major overthrust (heavy line). Curved fault traces result from folding and unequal erosion. After Keith.

The solution of the problem as to what becomes of a fault plane with depth will throw much light on earth structure, and evidence should be closely watched for.

After fracture planes are formed, they may be faulted or folded. Folded thrust-fault planes are described and figured by Keith in the Roan Mountain folio of the southern Appalachians [1] (Figs. 36, 37, and 38), and by Richards and

[1] Geol. Atlas U. S., Roan Mountain folio, No. 151, U. S. Geol. Survey, 1907.

Mansfield[1] in the Bannock overthrust in southeastern Idaho. When previously fractured rocks undergo conditions of flowage, the fractures tend to become obliterated.

Folded fault planes should not be confused with the curving of fault traces on the erosion surface, due to irregularities of topography.

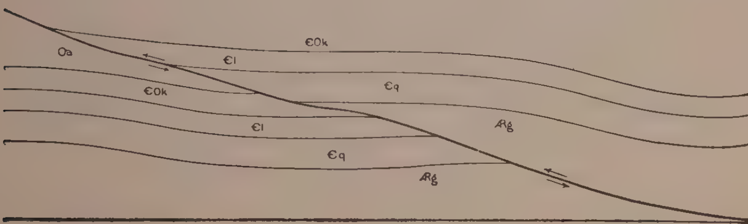

FIG. 37. Theoretical section showing supposed relations of beds in Fig. 38, after the major faulting but before the later folding and faulting. After Keith.

APPARENT AND REAL FAULT DISPLACEMENTS

One of the principal difficulties in the classification of faults lies in the confusion between the actual movement when viewed as a whole and the apparent movement seen in any two-dimensional section, whether this be vertical or horizontal.

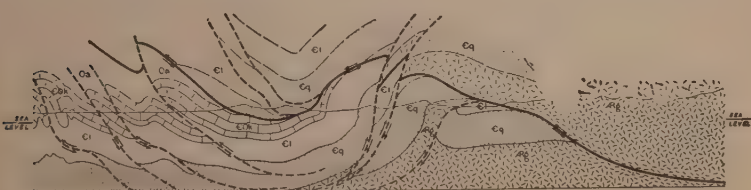

FIG. 38. Theoretical section across Buffalo Mountain and Limestone Cove, Tennessee. After Keith. Shows the character of the deformation and the relation of the younger faults to the older overthrust. Major overthrust, heavy continuous line; minor faults, broken heavy line; Oa, Athens shale and overlying beds; COk, Knox dolomite; Cl, Cambrian limestones and shales; Cq, Cambrian quartzites and slates; ARg, Archean granite and gneiss.

[1] Richards, R. W., and Mansfield, G. R., The Bannock overthrust: a major fault in southeastern Idaho and northeastern Utah: Jour. Geol., vol. 20, 1912, pp. 681–709.

Also the displacement as a whole may be the result of a series of movements not all in the same direction. Geologists are accustomed to plat their observations on a two-dimensional plan or section, and the names used for faults have been based mainly on the apparent displacements as they appear on these two-dimensional surfaces. Arrows are often used to indicate displacements, and the reader is likely to assume that these arrows indicate actual displacement, when as a matter of fact they indicate only the component of the displacement in the plane of the section. Consideration of the fault in three dimensions might require another name. For instance, a normal fault may be merely one part of a hinge fault or of a heave fault, another part of which will be described in cross section as a reverse fault. This confusion has perhaps arisen naturally from the fact that most of the common faults early recognized have, or were supposed to have, a dominant vertical element in their displacement. The recognition of horizontal movements and of hinge and pivotal faults came later. It is now clearly recognized that movements may range all the way from vertical to horizontal, and some effort is usually made to ascertain what the actual movement has been. Too often, however, the apparent displacement is assumed to be the real displacement, and terms are used which, while accurate for a given cross-section, may be quite misleading if taken to indicate the total displacement. One wonders whether on the whole it would not be better, for purposes of accurate description, to follow the method employed in the detailed mapping of faults in certain mining districts, where all the emphasis is placed on attitude of the fault and the actual displacement, without attempt to name the fault, — thus avoiding the confusion which might come from calling a unit structure a normal fault in one place, a reverse fault in another, a hinge fault in another, or a heave fault in another.

With alternative hypotheses open, how may the actual displacement be ascertained? It is frequently impossible to do

this, but there are certain ways in which the actual displacement has been worked out. These ways are as follows:

Striations may mark the direction of displacement. In some cases of repeated movements, later striations have crossed or destroyed earlier ones, perhaps formed in different directions. Repeated movements along fault planes, are especially common in faults due to cooling of igneous rocks.

The matching of the ends of broken dikes often makes it possible to determine the actual displacements of faults. This method has been used very effectively by the geologists of the Geological Survey of Scotland in determining both the direction and amount of the displacement of some of the large overthrust faults of the northwest Highlands of Scotland. Careful petrographic discriminations of these dikes and their uniformity and trend have aided greatly in tracing the dikes individually and in sets.

The matching of displaced ore-bearing veins has often indicated the actual displacement of faults. Probably in few other cases have the displacements of faults in three dimensions been considered so carefully as they have in many mining camps.

Displacement of topographic features at the surface may indicate actual displacement.

Nature of Fault Movements

Fault movements may be sudden, slow, or recurrent. Where the movements are recurrent they may or may not be in the same direction. Sometimes later fault striations and grooves are found crossing earlier ones. Later movements tend to obscure evidences of earlier movements, with the result that about all the observer can do is to ascertain the net displacement without knowing what combination of movements has really contributed to this result. In other cases sedimentary or igneous deposits uncomformably overlying an old break may themselves become faulted by a renewal of movement along the old fault surface.[1] Increasing knowledge of

[1] Irving, R. D., and Chamberlin, T. C., Observations on the junction between the Eastern sandstone and the Keweenaw series on Keweenaw Point, Lake Superior: Bull. 23, U. S. Geol. Survey, 1885, pp. 118–119.

the details of faults, especially where they exist in complex systems, is leading to the recognition of the fact that the actual movement is often far from simple.

In highly brittle rocks faulting may occur without much previous distortion of the rock mass, but ordinarily there is evidence of a previous accumulation of strains, represented in bending or folding and in differential migrational movements of the mass, in any direction from vertical to horizontal. Movements along the San Andreas fault of California (responsible for great earthquakes) have been definitely connected with the slow migration of the region as a whole as determined by geodetic measurements. These movements are known to vary in velocity laterally and are supposed to vary vertically. After a period of differential drag, during which rocks are bent, both within the elastic limit and beyond, the strains are relieved by a sudden fault movement or rebound (see pp. 264–266). This tends to restore the configuration existing before distortion. Relative displacement of parts of the region may be actually greater before than after the faulting. The rebound hypothesis may have broad application to faults, although it has been studied in only a few places. If so, it will considerably modify the prevailing geologic conception that the fault displacement itself is the culmination of regional distortion. From the rebound point of view it is only a localized displacement representing a residual of what was earlier a greater displacement of adjacent parts of the region.

Another interesting case of faults and joints related to migrational movements is cited by Brouwer in the Dutch East Indies, previously referred to.

Shortening and Elongation of the Earth's Crust by Faulting

Detailed studies of actual fault displacements are so few and far between that little can be said as to the actual elongation or shortening of large parts of the faulted earth's crust. Until

this is done, it is perhaps premature to consider general questions like the shortening or elongation of the earth's crust in a faulted area. Attempts have been made which suggest some of the following tentative considerations:

The displacements in normal faults have been assumed to be dominantly radial with regard to the globe, and as the dips of the fault planes are seldom exactly vertical, the downward movements require extension of the horizontal surface. Thrust or reverse faults have been supposed in general to represent tangential shortening, with subsidiary vertical displacement.

In view of the difficulties, already referred to, of determining locally whether a fault represents tension or compression in three dimensions, it is obviously impossible to answer the question for large areas as to the quantitative effect of faulting in extension or shortening of the earth's surface. For a given area tension faults at the surface may be much more numerous than thrust faults, yet the lengthening of the surface represented by the tension faults may be of less amount than the shortening of the surface by one thrust fault. The dip of a thrust-fault plane is usually low, that of a normal fault plane, high. An average from the United States Geological Survey folios gives a dip of 36° for reverse fault planes and 78° for normal fault planes. A displacement of a foot on the dip of the thrust-fault plane means nearly a foot of horizontal shortening; a displacement of a foot on the normal fault means but a few inches of horizontal lengthening. A single thrust plane of low dip, then, may accomplish a horizontal shortening which would require a large number of normal faults to compensate. Actual displacements of tens of miles have been observed in low-angle faults while displacements in high-angle faults seldom reach a mile.

If the crust as a whole has been shortened by mountain folds, it might appear that thrust faults are the dominant structure, and that normal faults are ultimately subsidiary phenomena.

Geologic history seems to point to alternations of great compressive and relaxational movements. During a period of mountain-making compressive stresses develop, resulting in tangential deformation in a comparatively short space of time, with subsidiary radial deformation in areas of uplift. During the succeeding period of quiescence it may be supposed that the action of gravity on uplifted areas may develop normal faults, which partially compensate for the earlier shortening.

The extension of areas caused by normal faults due to the cooling of igneous rocks or the drying and settling of sediments is commensurate with the shrinkage of these rocks during these processes; such faults may cause no real extension of the earth's surface.

It is not certain that all thrust faults involve tangential shortening of the earth's surface. Davis [1] has suggested that because advancing thrusts leave no cavity behind, and because friction may prevent a shove of the whole earth's crust, the overthrust material is supplied by an oblique extrusion of subcrustal matter, in the form of a wedge, which would actually extend the earth's surface. He suggests also that this would involve the "underdrag" of the crustal area behind, perhaps breaking by normal faulting. He cites one of the Basin ranges as a possible illustration of this process.

There have been some attempts to calculate the lengthening or shortening of an area on the assumption that the displacements shown in cross-sections are the real displacements, without taking into account the probability of displacement in the third dimension. One of the few attempts to consider the problem in three dimensions is that of Emmons and Garrey, who have estimated the actual extension by faulting of the Bullfrog district of Nevada.[2] They show that the apparent extensions in individual cross-sections are greater than the real extensions because there has been much movement in

[1] Davis, W. M., Bull. Geol. Soc. Am., 1923 (to be published).

[2] Ransome, F. L., Emmons, W. H., and Garrey, G. H., Geology and ore deposits of the Bullfrog district, Nevada: Bull. 407, U. S. Geol. Survey, 1910, p. 88.

directions inclined to the plane of the cross-section shown by striations on fault surfaces. From somewhat careful quantitative study they conclude that the apparent extension should be reduced by at least one-third to approximate the real extension of the area.

In short, the net effect of faults in crustal shortening is still a largely speculative field.

Evidences of Faulting

Few faults are so obvious that they make themselves immediately known to the observer. Faults of hundreds of feet of displacement which are actually exposed at the surface and in underground openings have escaped detection for years, even where the ground has been mapped by skilled observers. For considerable parts of the earth faulting is perhaps the exception, rather than the rule, as the controlling or even the most important feature of structural mapping. The geologist here carries his mapping and interpretation forward on the assumption of more or less normal sequence of beds, and of distribution and relationships primarily determined by folding. It is only where these assumptions do not suffice to explain the observed facts that his attention is focused on the possible existence of faults. In a much faulted area the initial assumptions are likely to be made in reverse order.

A fault is seldom completely observable in all its aspects; it usually reveals itself slowly, through bits of evidence which require careful inductive inferences and correlation. No one fault may exhibit all the different kinds of fault evidences, and no one line of evidence may be conclusive. Some of the kinds of evidence ordinarily used are as follows:

(1) A fracture along which there is polishing or slickensides or striations showing the movement of one side with reference to the other. These are often found also along joints where the movement has been too small to warrant classifying the structure as a fault.

(2) The presence of gouge in and along the fracture indicating the grinding up of the rock. Sometimes the gouge is merely ground-up rock. In other cases it is ground-up rock which has been leached and altered to clayey substance. It is usually soft, highly charged with water, and in underground openings causes falls and caves which may call attention to its existence. In walking through a mine the presence of gouge-falls on the floor may call attention to the existence of a gouge or fault in the roof. Such places of weakness are often timbered so that in the underground mapping of ground so hidden some effort should be made to find out what exists behind the timbers. The gouge may take on a cleavage along S-shaped surfaces, which may indicate the direction of the movement under the principles outlined on page 185.

(3) Brecciation of the rock along a fissure may also suggest movement.

(4) The presence of many parallel fractures constituting a " shear zone " is a common evidence of faulting.

(5) Abrupt drag or folding of beds may suggest the proximity of a fault. The direction of the displacement may also be so inferred.

(6) Where the movement has been considerable there may be large grooves or furrows on the fault walls, sometimes called *mullion structure*. This structure indicates the direction of the displacement. The grooves may be several feet from crest to crest. Fine examples of these structures may be seen along the great thrust planes of the Highlands of Scotland, in the Butte granite, and perhaps in the Rodadero fault of Cuzco, Peru.[1] Where very small they can be used only with reservation, as indication of the major direction of faulting, because a minute structure of this kind may be the result of local movement or slumping.

(7) It is sometimes possible to observe directly the fact

[1] Gregory, Herbert E., The Rodadero (Cuzco, Peru),—A fault plane of unusual aspect: Am. Jour. Sci. vol. 37, 1914, p. 289.

that beds or dikes or other original structures do not match up on opposite sides of the fracture, thus showing that there has been some displacement.

(8) Direct evidences of faulting of the kind above named are satisfactory criteria where found, but in a vast number

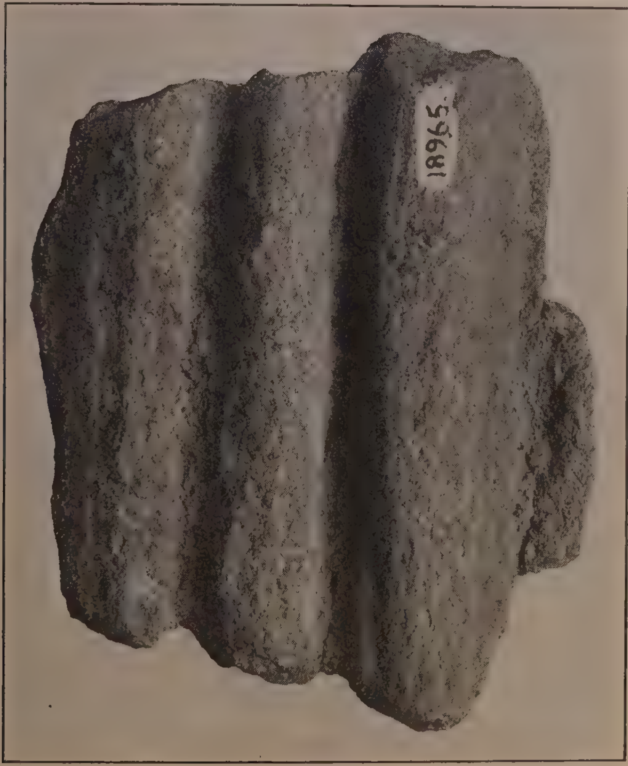

FIG. 39. Mullion structure.

of cases the geologist is not fortunate enough to find them. Many faults are mapped which are never seen, because they are the only structures which can explain the space relations of beds or other preëxisting structures. Beds are repeated or eliminated. They are known to be out of their normal sequence and contiguous to stratigraphically separated beds.

Normal stratigraphic or intrusive relations, or even folding, will not suffice to explain their relative positions. When we consider the possible variety of fault directions, angles, and kinds, and the varied and complex original structures which

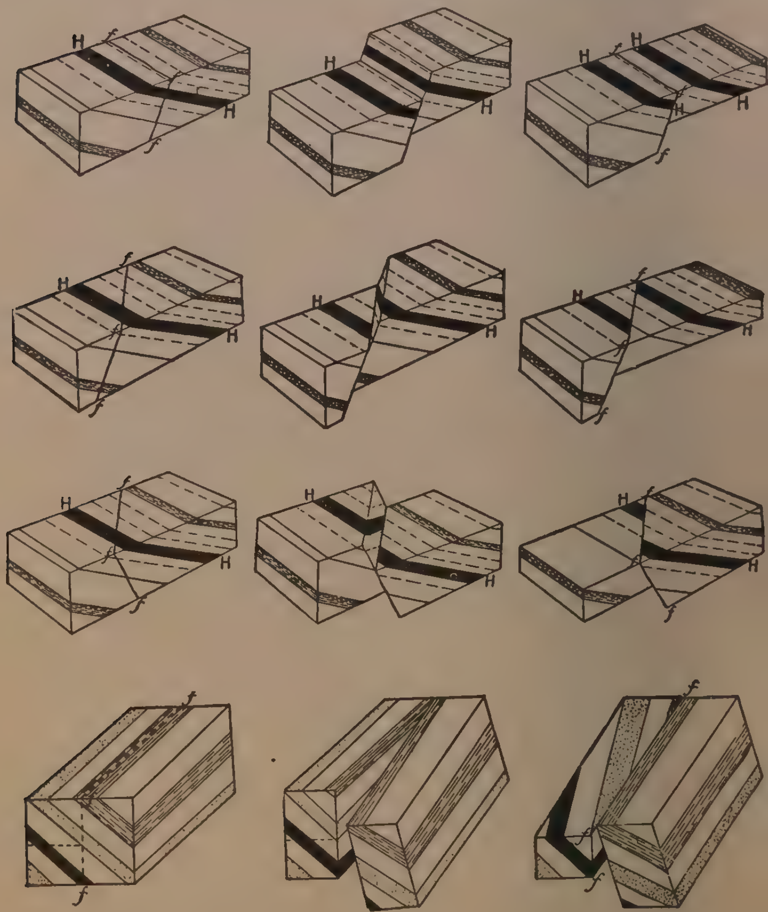

FIG. 40. Illustrating displacements of bedding by faulting. After Chamberlin and Salisbury. These are only a few of the many possible relationships that might be shown.

they cut, and especially when we superpose on this an irregular erosion surface, it is clear that the irregularities of distribution which may be interpreted as due to faulting may be almost infinite in number. It is impossible to set forth all the various combinations. Most geologists get a real understanding of the problem only by field experience.

(9) On the surface there may be an actual fault scarp formed directly by the faulting. This, however, is comparatively rare. More often there is a scarp formed by erosion of a softer rock brought into juxtaposition with a harder rock by the faulting.

Fig. 41. Fault scarp developed during the earthquake of 1899, Yakutat Bay region of Alaska. After Martin.

A fault scarp resulting from recent displacement accompanied by earthquakes is illustrated in Fig. 41.

A more common evidence of recent faulting is a shallow trough along the fault trace.

Among the best known instances of faults still represented by the original escarpments are the Hurricane fault separating the Wasatch Mountains from the Great Basin, and the faults

of the Great Basin ranges, which were originally classified by Gilbert [1] as an example of the block type of mountains. The criteria used by Gilbert for the recognition and delineation of fault scarps of the Great Basin are (*a*) their steepness, (*b*) their association with shear zones and displacements of beds, (*c*) displacement of plateau level, as in the Hurricane fault of Utah, (*d*) the fact that the scarps may not converge toward the ends of the mountains as they would if they were normal erosion scarps, (*e*) the existence of triangular facets across the ends of ridges as though the ridges had been sliced off, (*f*) the recent displacement of alluvial fans, lake beds, and other surface features, indicating that the faulting has been going on to very recent times and has not had time to be masked by erosion. Spurr [2] questions these criteria for the surface delineation of faults, or rather, the degree of emphasis to be placed on them. He calls attention to the fact that erosion has been conspicuously effective in producing the present topographic features of some of the Great Basin ranges, that anticlines and synclines play an important part, and that the recognized faults in these ranges are often quite independent of the topographic features. The student may study maps of these ranges to advantage, keeping in mind the criteria above cited.

In districts of older deformation, like the southern Appalachians, erosion has had a longer time to develop the topographic features, with the result that original fault scarps are practically non-existent. The effect of faults on the topography is due to their bringing into contact rocks of unequal hardness, thus permitting differential erosion. The flat and curved attitudes of the fault planes here also tend to make them less conspicuous in the topography.

Thrust faults are not likely to produce steep vertical es-

[1] Gilbert, G. K., Report on the geology of portions of Nevada, Utah, California, and Arizona, examined in the years 1871 and 1872: U. S. Geog. Surveys W. 100th Mer., vol. 3, 1875, pp. 17–187.

[2] Spurr, J. E., Origin and structure of the Basin Ranges: Bull. Geol. Soc. Am., vol. 12, 1901, pp. 264–266.

carpments, either before or after erosion. By pushing forward successive slices of rock they tend to cause linear features in the topography, and yet these features are not different from those which might have been produced by folding, and probably they would not be identified as related to thrust faulting unless the thrust faulting had been otherwise proved.

In regions of vertical faults, especially in flat-lying beds and non-glaciated areas, the lines of the faults are very likely to be marked by drainage channels which have developed along these planes of weakness. All faults are not marked by drainage lines, nor do all drainage lines mark faults. (See pp. 62–63).

One of the most fully studied cases of the surface expression of a fault with horizontal displacement is that of the fault which caused the California earthquake of 1906. "At the surface the cracks had great variety of expression. Some were barely perceptible as partings; others gaped so widely that one might look down them several yards. Some were mere pullings apart; others disclosed small differential movements of the nature of faulting. Some were solitary; others, especially those exhibiting faulting, were in groups."[1] Where the fault crossed a spur or shoulder of a mountain a scarp appears. Small basins or ponds, many having no outlets and some containing saline water, are frequently found at the base of small scarps. Trough-like depressions appear on both sides, also bounded by scarps. Small knolls or sharp little ridges are common at the fault line and these are bounded on one side by a softened scarp and separated from the normal slope of the valley side by a line of depression. Other effects of this fault are slides of earth or rock from the hill-slopes. Finally, there are many conspicuous dislocations of the works of man.[2]

[1] Gilbert, G. K., Bull. 324, U. S. Geol. Survey, 1907, p. 7.

[2] Lawson, A. C., Preliminary report of the State Earthquake Investigation Commission, 1906, and final report, 1908.

CHAPTER VI

BRECCIAS AND AUTOCLASTICS

When rocks are broken into irregular angular fragments they are called *breccias* or *autoclastics* ("self-broken" rocks). They may or may not be cemented. Breccias are common accompaniments of jointing, faulting, and folding. Even deformation mainly by rock flowage may show brecciation in brittle layers. Breccias form in landslides and slips. Newly deposited sediments may be brecciated by wave or tidal action, scour or slip, or the breaking of case-hardened surfaces (see p. 207). Breccias are common in clays or limey muds which have been temporarily exposed to air. Volcanic explosions and the flowage of magmas produce breccias. Expansion and contraction of rocks, due to changes in temperature or in dehydration, or in mineralogical and chemical composition, form breccias. Especially does solution lessen volume, causing "solution," "slump" or "founder" breccias, characteristic of certain mineral deposits. In fact, all the forces which deform rocks may be effective in forming breccias under some conditions.

A great variety of names and classifications of breccias has been proposed. Breccias are formed under so many different conditions that the name used for any one depends largely on what particular feature the investigator of the moment wishes to emphasize. It is often desirable to qualify the term breccia by some adjective descriptive of origin or the type of rock. It is doubtful if any general classification, sufficiently comprehensive to meet all conditions, and logical in its discrimination between processes of origin and kinds of rock, can be agreed to by geologists. For a general summary of this sub-

ject the student is referred to Norton's "Classification of Breccias."[1]

The principal source of difficulty in naming is that breccias formed by diastrophic movements, sometimes grouped as

FIG. 42. Friction breccia in quartzite, showing separation of fragments by vein quartz, Baraboo, Wis.

[1] Norton, W. H., A classification of breccias: Jour. Geol., vol. 25, 1917, p. 160.

friction breccias, may be identical with, or similar to, or grade into, or have common features of origin with, breccias formed by volcanic explosion and deposited in beds of tuff or agglomerate; with rocks formed under conditions of primary deposition, sometimes known as intraformational conglomerates; with basal conglomerates; or with surface rubble and talus resulting from weathering and erosion.

Thus: Surface weathering of a rock may break it up into small fragments, which, if in place, and especially if cemented, might be called breccia. Preëxisting fractures aid and localize this process. There is a general tendency for the residual fragments to be rounded, but in some kinds of rocks, especially slates and quartzites, the fragments remain angular. A surficial breccia or rubble of this kind may remain in place and become the basal conglomerate of an overlying sedimentary deposit, either subaqueous or subaerial. It is then usually known as a conglomerate, but it grades below into brecciated rocks so closely associated with the primary rock that part of it may be called breccia. There are many unconformable surfaces, particularly in the pre-Cambrian, where it is very difficult to draw the line between the older brecciated surface of the lower formation and the conglomeratic base of the upper formation. The intermediate rock in some such cases has been called a *recomposed* rock.

Similarly, the heating, cooling, and crystallization of magmas cause many faults and joints which locally pass into breccias; and it is sometimes difficult to be sure, unless all the field relations are clear, whether such a rock should be called a breccia, implying ordinary breaking of the solid rock in place, or whether it should be called a tuff, implying some movement through the air. The hardened surface of a flow may be broken and crushed, forming a *flow breccia* which it is sometimes difficult to distinguish from a tuff or from a friction breccia. Also magmas under some conditions invade the country rock by a process of stoping, with the result that

angular fragments become imbedded in an igneous matrix. The resulting rock is truly a breccia, but one of special origin.

When heterogeneous beds are subjected to rock flowage it commonly happens that the more brittle layers and minerals brecciate, while the softer ones flow. Broken fragments are cemented by the flowing softer material. Parts of such a rock mass are called breccias, while other parts may become schists. On a minor scale, and especially on a microscopical scale, rock flowage is nearly always accompanied by this kind of brecciation in the more brittle layers or minerals, but usually the term breccia is not applied unless a considerable part of the mass shows brecciation on a macroscopical scale.

In short, rocks may be broken into breccia-like forms under a great variety of conditions and by a great variety of processes, and there are all gradations between these conditions and processes. There are, therefore, no hard and fast lines which separate classes according to origin. Furthermore, even where there are fairly distinctive differences in origin, the resulting rocks may have much the same physical characteristics. It is necessary to carry into the field a wide comprehension of the various possibilities to avoid misinterpretation of breccia-like structures. It has been the writer's experience that this is one of the most common sources of mistake. Especially in regions which have been later deformed by rock flowage, more or less obscuring the earlier features, it is necessary to be on one's guard in the interpretation of these elusive structures. A geologist who has been working in an area where there is clear evidence of autoclastics may carry preconceived notions of the abundance of that kind of rock into another area, and fail to recognize there the existence of a conglomerate of great stratigraphic significance. Another, who has dealt principally with conglomerates and seen little of autoclastics, may assume a rock to be a conglomerate without sufficiently considering the possibilities of its being autoclastic.

Notable controversies in pre-Cambrian stratigraphy and structure have been due to differing interpretations of this class

of rocks. Owing to the bias of preconceived opinions various interpretations of the origin of the fragmental rocks at the base of the Huronian series in the "Original Huronian" district north of Lake Huron for a long time caused controversy, and delayed a true understanding of the geology of this important area. The base of the Algonkian in northern Minnesota is made up of fragments of the underlying Archean basalts, porphyries, and granites, — which, over considerable areas, still retain their angular form, and evidently have been but little worn and transported. Lying unconformably beneath them, and associated with the underlying igneous rocks, are various autoclastic and tuffaceous forms so similar in characteristics to the basal conglomerate that, even with the application of the most careful criteria, it is frequently difficult or impossible to tell whether a given exposure of rock should be mapped as Algonkian or Archean. In the progress of the mapping the interpretation of the geology has changed from time to time according as these rocks came to be better known.

When, after all the criteria have been used, there is still uncertainty as to origin, this should be clearly indicated, in order to keep the subject open for further investigation.

Criteria for Identifying Breccias of Various Origins and Conglomerates

Friction breccias and conglomerates. (*a*) The fragments in the friction breccia are usually more angular than those of the conglomerate, but to this there are exceptions. The pebbles of a conglomerate, when made up of some types of rock, or when little transported, may also be angular. On the other hand the fragments of a friction breccia may be remarkably rounded by weathering before cementation, or by attrition during later rock flowage. (*b*) The fragments of a friction breccia are likely to be more homogeneous in character; ordinarily they are of one kind of rock. Conglomerates may have several kinds of fragments coming from different sources. However, many conglomerates are made up domi-

nantly of one kind of fragments, hence this criterion is also inconclusive. (*c*) The cement of a friction breccia is likely to be vein material. This is one of the safest criteria. (*d*) A friction breccia may be developed in zones passing through bedding, a position not held by conglomerates.

No one of the above distinctions is decisive. Collectively they may be so, but not in all cases.

Criteria for the separation of friction breccias from intraformational conglomerates and breccias, supposed to be the result of forces acting during or soon after primary deposition, are about the same as those above cited. The intraformational conglomerates are really structures representing an intermediate stage between conglomerate and breccia and might be called either breccia or conglomerate, depending on whether the structural or the stratigraphic aspects of the rock are to be emphasized.

Tuffs and volcanic breccias. Means of identification of tuffs and volcanic breccias are often found (1) in the lithologic homogeneity of their fragments, (2) in the possession of volcanic textures, such as amygdules peripherally arranged, (3) in the cementation of their fragments by volcanic dust or rock, (4) in the angularity of the fragments, (5) in the irregularity of size of fragments, (6) in their occurrence in beds between volcanic flows, and (7) in other field relations. None of these criteria is decisive. Tuffs formed under water are distinguished only with very great difficulty from water-deposited clastics resulting from erosion. The fragmental rocks associated with the immense basaltic flows of Ontario and the Lake Superior region well illustrate the difficulties of this discrimination.[1]

Porphyritic and amygdaloidal textures. Volcanic rocks having amygdaloidal fillings or porphyritic textures may simulate conglomerate, especially on weathered surfaces or in schistose rocks. The mineralogic character of the pseudopebbles will usually serve to fix their origin.

[1] Van Hise, C. R., and Leith, C. K., Geology of the Lake Superior region: Mon. 52, U. S. Geol. Survey, 1911, pp. 118–143.

Cementation of Breccias

Some geologists use the term breccia only where the broken fragments have been cemented into a coherent mass, and for the unconsolidated fragments use other terms like rubble or talus. More generally, however, the term is used alike for both consolidated and unconsolidated masses of broken rock. Almost any kind of rock or mineral may serve as cement, — igneous and sedimentary rocks, calcite, dolomite, quartz, iron oxide, etc. Of all the cements named, possibly vein quartz is the most abundant and widespread, particularly in the older rocks. It is a well-known fact of metamorphic geology that quartz tends to replace calcite and dolomite cements in the continued metamorphism of rocks.

Where the rock is partially cemented it is seen that the minerals coat the fragments and work outward into the center of the cavities. Quartz crystals, particularly, project outward into the center. Where the rock is completely cemented the space has been entirely filled by the meeting of these crystals. It is a question whether the growing of crystals is stopped with the closing of the original cavity, — for the reason that some breccias show fragments completely separated from one another and imbedded in quartz, in such a fashion as to indicate that the growing quartz has exerted pressure to separate the fragments. If the cementing quartz of the typical breccias of the Baraboo quartzite were all to be taken away, the fragments of quartzite would slump to regain contact with one another (Fig. 42). This phenomenon is supposed to be the same as the widening of joints by the force of growing crystals, discussed on pages 60–62.

CHAPTER VII

FLOW CLEAVAGE AND FRACTURE CLEAVAGE

Flow Cleavage Described

Flow cleavage is a structure commonly resulting from flowage of hard rocks. It is a capacity to part along parallel surfaces determined by the parallel arrangement of the longer axes of unequidimensional mineral particles and by the parallel arrangement of mineral cleavage in certain of the unit mineral particles. Flow cleavage is characterized by platy and columnar minerals of comparatively few kinds which are well adapted to conditions of rock flowage. The nature of this adaptation presents an interesting problem referred to later (p. 144).

Flow cleavage is not an inevitable consequence of rock flowage. Materials like soft muds and sands may flow without taking on a cleavage. Even hard rocks like limestone or marble may flow without developing it. A rock deformed by flowage also may assume a rudely banded structure, characteristic of certain gneisses, and still not have flow cleavage.

Other names for flow cleavage are *schistosity* and *slaty cleavage*. The rocks exhibiting these structures are called *schists* and *slates*. Slaty cleavage is ordinarily distinguished from schistosity in straightness of cleavage planes and fineness of grain, but the distinction between the two structures is not sharp — one grades into the other. Cleavage, even of slate, is never perfectly uniform or straight, but small slivers or leaves are left on cleavage surfaces. As the names implies, slaty cleavage is best developed in fine-grained argillaceous rocks, but fine-grained igneous rocks like basalt or rhyolite may also take on this structure. Schistosity ordinarily develops in rocks of coarser grain. There are exceptions to this general-

ization, for a good slaty cleavage may form even in quartzites, and under conditions of intense metamorphism the slaty cleavage characteristic of fine-grained rocks may by recrystallization take on the coarser structure of schistosity.

Gneissic structure is in part included under flow cleavage but in part it is something else, hence this structure is treated independently. The "crystalloblastic" structure is another designation for flow cleavage (see pp. 141–143).

Flow cleavage is most typically characterized by the presence of mica, chlorite, and hornblende in parallel dimensional arrangement. These minerals, together with quartz and feldspar, make up all but a very small percentage of schistose or cleavable rocks. To make the discussion concrete, therefore, cleavage will be discussed principally in relation to these minerals. The technical reader will at once think of qualifications and additions necessary where other minerals are considered, but in the writer's judgment these do not essentially affect conclusions based on the study of a few of the principal schist-forming minerals.

One of the peculiar features of a cleavable rock is the uniformity in shape of the grains of each of the characteristic minerals, determined by the crystal habit. The average ratio of the greatest to the mean dimension of a mica plate is about 10:1, of hornblende 4:1, and of quartz and feldspar 1.5:1. These ratios are the same whether the rock cleavage is good or poor. In other words, the better rock cleavage does not necessarily mean a greater drawing-out or elongation of mineral particles.

When, in the laboratory, crystals are allowed to develop under stress, they elongate in the plane of easiest relief, — supposedly regardless of habit, but this is not certain, because the experiments have been conducted principally with isometric crystals.[1] Also crystals not under conditions of growth have been elongated by pressure alone, again more or less

[1] Becker, G. F., and Day, A. L., Linear force of growing crystals: Proc. Wash. Acad. Sci., vol. 7, 1905, pp. 283–288.

regardless of habit. But, notwithstanding these experimental results, the minerals in schists are known by observation to have a shape ordinarily determined by crystal habit alone. The difference between a schist with poor cleavage and one with good cleavage is not so much that the particles of one have been elongated more than the particles of the other, but that one has more of the kinds of minerals which by crystal habit are elongated.

As a necessary consequence of the parallel dimensional arrangement of mica flakes and hornblende crystals, their mineral cleavages also take on some uniformity of arrangement. Mica flakes, for instance, lying dimensionally parallel in a schist, have their mineral cleavages in the plane of the two greater dimensional axes, that is, in the plane of rock cleavage. Hornblende crystals lie with their long dimensional axes parallel; the mean or least dimensional axes of hornblende crystals, being so nearly of the same length, may not be parallel. The two cleavages of hornblende are parallel to the major dimensional axis but are inclined to the minor dimensional axes. Thus the hornblende cleavages in the schistose rocks are parallel to an axis, but not to a plane. The feldspar habit does not give such great dimensional differences. Most of the feldspars in schists show only a slight tendency to assume elongated or tabular shapes due to crystal habit. Their dimensional arrangement is more or less independent of crystallographic arrangement, and therefore there is only a slight tendency toward parallelism of the feldspar cleavages. In exceptional instances the feldspar particles exhibit both dimensional and crystallographic parallelism, and the feldspar cleavages have an effect on the rock cleavage. The good basal cleavage stands almost at right angles to the greater dimensions of the crystal, and hence to the plane of rock cleavage, to which the greater dimensions of the crystals are parallel. As the greater diameters of the crystals are parallel only to the plane of rock cleavage and not to each other, the basal cleavages of the different crystals, while always at right angles to the

plane of the greatest and mean diameters, may be at any angle to one another. Quartz does not have a good mineral cleavage, and even if it had there would be little parallelism in the cleavage of the different particles because the dimensional habit is so stumpy that there is little tendency to parallelism of crystallographic axes.

A schistose rock cleaves either between the mineral particles, following the plane of their greatest and mean dimensional axes, or within the mineral particles along their cleavage planes. The first is known as inter-mineral cleavage, and is a capacity to part determined by the dimensional arrangement of mineral particles; the second may be called inter-molecular cleavage, and is related to the ultimate molecular structure of the crystals. Ordinarily when a rock is cleaved the two surfaces show the glistening faces of hornblende or mica or of other minerals of this type, indicating that the break has followed the mineral cleavages. The parting here has obviously been easier than between the mineral particles. In places where mica and hornblende are not abundant, the cleaved surfaces of the rock show quartz and feldspar, indicating that the breaking has been principally of the inter-mineral type.

Whatever the relative importance of inter-mineral and inter-molecular cleavage, it should be remembered that most mineral particles in cleavable rocks are dimensionally arranged, and that this dimensional arrangement involves parallelism of the mineral cleavages for only part of the minerals. Therefore the conclusion is justified that the dimensional parallelism of mineral particles is the controlling factor in rock cleavage, and that to this control is due the mutual parallelism of mineral cleavages of mica or hornblende. Nevertheless it may to some extent be true that the cleavages of these minerals have some influence on their elongation, and therefore on their arrangement. As the dimensions of the minerals of schists are controlled by mineral habit, this becomes an important factor in the structure of schists.

In addition to the dominant feature of parallel arrangement above described there is an endless variety of pressure effects, expressed in the strained and broken condition of the mineral particles. The straining of a mineral, particularly quartz, modifies the optical properties, with the result that under the microscope, between crossed nicols, the crystal shows dark shadows which move across it as the stage is rotated, — owing to the directions of extinction differing from point to point. These are known as *strain shadows*. Quartz sometimes shows rows of fluid-filled pores, marking directions of shearing planes, which may be traced through contiguous crystal grains. Micas may yield by shearing movement causing lamellar twinning, parallel to gliding planes, or they may show a bending of crystals. Plagioclase may show a secondary twinning. Orthoclase may develop a *microcline* structure. Pyroxenes, feldspars, and olivine may show *schiller* structures, which are explained by development of cavities, subsequently filled, along easy solution planes, which may be planes of gliding. The breaking of crystals is much more conspicuous in quartz than in feldspar because of its more brittle nature. The crystals may be broken along parallel planes and the parts relatively displaced, as the result of slicing. All stages of cracking and granulation are to be observed, although in some schists these are obscured because recrystallization has healed the fractures. The textures determined by fracture are known in general as *cataclastic* structures. Granulation may be confined to the periphery of grains or groups of grains, leaving the cores unbroken. This is sometimes called the *mortar* structure, or the *augen* structure. The complete granulation of constituents is sometimes called the *mylonitic* structure, although this term in actual usage covers also schistose structures in which evidence of recrystallization is important.

Manner in Which the Parallel Arrangement of Minerals is Brought About

The arrangement of the mineral constituents of a cleavable rock is the result of the differential pressure which caused the rock to flow. It remains to describe more in detail just what the processes of rock flowage are.

Recrystallization. A study of cleavable rocks shows that much of the hornblende and mica, minerals which are responsible for some of the best rock cleavage, is of entirely new generation in the secondary rock. A shale or mud may have little mica; a phyllite, its altered equivalent, may have as high as 50%, by weight, of mica. Chemical analysis shows that such changes may occur in some instances with little addition or subtraction of materials. A correct inference is that the new minerals of the horneblende and mica types have developed principally from the recrystallization of substances already in the rock mass. Even where there is quantitative evidence that substances have been introduced or extracted, the mass has still been recrystallized. Since hornblende and mica are the common minerals producing the best rock cleavage, it must be concluded that recrystallization is the important process in the development of parallelism of the mineral constituents, and thus of rock cleavage.

Corroborative evidence of the importance of recrystallization is the fact that, in general, fractures and other strain effects in the minerals of a cleavable rock are not on a scale to be expected if the parallelism had been brought about entirely or largely by mechanical processes. It may be inferred, then, that some constructive process, which may be called generally recrystallization, has been at work.

Most of the mineral particles in the cleavable rocks are individually larger than the particles in the same rocks before flowage occurred. For instance, the gradation of a slate to a phyllite means an increase in the size of the grains. Recrystallization is the constructive process which has accomplished this result.

The cleavable rock is likely to show a greater uniformity in size and shape of the grains of the same mineral as compared with the non-schistose rock, and again recrystallization explains the phenomenon.

Much detailed microscopical evidence might be cited, — such as dove-tailing of quartz individuals in quartz bands, the feathering out of mica plates against adjacent mineral surfaces, the lack of bending and breaking of hornblende needles by

FIG. 43. Photomicrograph of micaceous and quartzose schist showing recrystallized quartz. From Hoosac, Mass. The view illustrates in detail the relation of recrystallized quartz grains to recrystallized mica flakes. The mica flakes for the most part separate different quartz individuals, but they may be seen to bound two or more individuals and to project well into them. It is not probable that such a relation could be brought about by granulation, slicing, or gliding, and it seems best explained by recrystallization. Bull. 239, U. S. Geol. Survey.

mutual interference, the segregation of minerals into bands, — to show that the parallelism could not have been produced by mechanical adjustment alone, but must have been aided by the chemical and mineralogical changes involved in recrystallization. (See Figs. 43 and 44.)

Granulation and rotation of original particles. Recrystallization is not the only process instrumental in the production of rock cleavage. The quartz and feldspar in the cleavable rock may be largely original quartz and feldspar; some of the mica and hornblende also may be original. Parallelism may be partly due to rotation from original random positions. This process may be aided by granulation and

FIG. 44. Photomicrograph of micaceous schist from Hoosac tunnel. The micas, which are entirely new developments by recrystallization, lie in flat plates with their greater diameters roughly parallel. Each individual exhibits several twinning lamellae. It will be noted that, while there is apparently a bending and irregularity in the mica plates, the individuals are for the most part not deformed, and the impression of irregularity is caused by the individuals feathering out against one another at low angles. This sort of arrangement is frequently seen about rigid particles which have acted as units during deformation, indicating that the arrangement is due to differing stress conditions at different places. Bull. 239, U. S. Geol. Survey.

slicing of the original mineral particles. Broken, unequidimensional mineral fragments are often strewn out in such a

manner that their longer dimensions lie approximately parallel. Evidence of rotation is seen principally in the quartz and feldspar, which have not much effect in producing rock cleavage. It is concluded, then, that the rotation of original particles, diversely oriented, to a parallel position is a minor factor, quite subordinate to the dominant process of recrystallization. (See Figs. 45 and 46.)

In the incipient stages of rock flowage the larger and more brittle particles are granulated and elongated. At the same time recrystallization, beginning on the finer particles, builds up new minerals. In the intermediate and advanced stages it gradually dominates over granulation and ultimately obliterates any evidence of it. It may be inferred that granulation aids recrystallization in that it grinds the particles into small pieces and affords greater surface upon which the chemical process may act.

English and Scotch geologists designate gran-

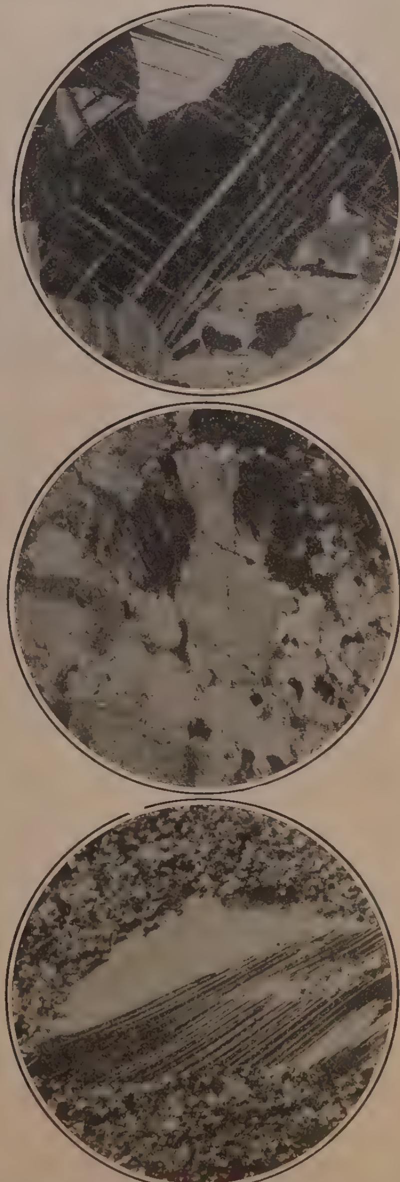

FIG. 45. Photomicrographs showing the progressive granulation of the Morin anorthosite under the influence of pressure. After Adams.

ulated structure in part as *mylonite*, which is composed mostly of minute fragments and particles, with now and again larger fragments set in a streaky groundmass of crushed materials. These have been especially described in connection with the great over-thrust faults of the Highlands of Scotland, where they are closely associated with and grade into crystalline schistose rocks, formed by the same movement. The writer has examined some of these mylonites and from his point of view all come under the general heading of schists caused by rock flowage, mylonites meaning merely parts of the mass

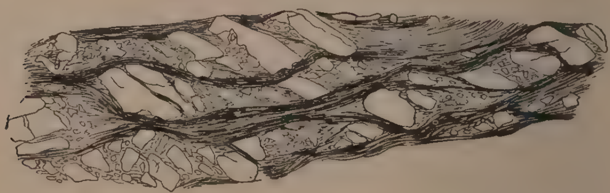

FIG. 46. Sliced feldspars in micaceous and chloritic schist from southern Applachians. Bull. 239, U. S. Geol. Survey.

where the evidence of breaking, which is always present in rock flowage, is somewhat more conspicuous.

In experimental deformation the conditions are not favorable for recrystallization and granulation is the important process.

Gliding and other processes. Slipping or twinning along the cleavage planes of minerals, called "gliding" — such as may be observed in calcite and ice crystals — has been cited as a possible cause of the elongation and parallel arrangement of mineral particles. This has been observed only in minerals of the calcite type, which are not important in cleavable rocks; and even in the calcite of schistose rocks gliding has been found to be subordinate to processes of recrystallization and granulation. In experimental deformation of marble it seems to play a greater part, because conditions of recrystallization are not present.

There is no evidence that the flattening of original mineral

particles to a dimensional parallelism, without regard to crystallographic arrangement, has played any important part in the production of rock cleavage; indeed, some of the facts already cited constitute decisive evidence to the contrary. Such is the evidence that hornblende and mica, essential minerals of schistose rocks, are in many cases, and perhaps in most cases, entirely new developments in the rock. Of the same nature is the evidence derived from the uniformity of dimensional characteristics of the particles of a given mineral species and the control of dimensions by crystal habit. The more cleavable rock is not made up of flatter particles of hornblende, mica, quartz, or feldspar than the less cleavable rock. But it certainly contains more particles of hornblende and mica than of quartz and feldspar; consequently it has more particles which are flat or elongate, which give it a better and smoother cleavage.

Cleavage in its Relations to Stresses

It has been shown that rock cleavage is determined by the parallelism of mineral constitutents and that this parallelism is developed by rock flowage, which implies differential pressures. It now remains to discuss the attitude of cleavage with reference to specific pressure conditions.

What experimental evidence there is indicates that in a non-rotational strain (see p. 18) mineral particles tend to arrange themselves with their longest dimensions normal to the direction of the pressure. There is practically no experimental evidence bearing on the arrangement of particles under rotational strain or shearing, so common in nature.

Wright [1] melted about 50 grams each of wollastonite, diopside, and anorthite, and plunged the melt into water, thereby forming a glass. Cubes were then cut from these glasses, heated to a viscous state at which crystallization first begins, and subjected to vertical pressure. Microscopic examination

[1] Wright, F. E., Schistosity by crystallization — a qualitative proof: Am. Jour. Sci., 4th ser., vol. 22, 1906, p. 226.

showed that the three minerals named had crystallized with their longer dimensional axes normal to the pressure.

Becker and Day [1] have shown that although crystals are able to grow in a given direction in spite of contracting forces, their growth in the plane perpendicular to the pressure is vastly greater, whether this be the normal direction of elongation due to habit or not. Ordinarily in schists the elongation of the crystal is that of its normal habit, indicating perhaps that the crystals favorably oriented to grow with normal habit have grown at the expense of those not favorably oriented.

Field observations have to do principally with the relation of cleavage to rock *strain* (see p. 15), which can be seen, and not to *stress*, which can not be seen and may only be inferred from the strain. After having proved the relation of cleavage to strain, the general relations of strain to stress may be considered. These two steps in the reasoning are separately discussed under (1) and (2) following.

(1) It seems self-evident that the longer dimensions of mineral particles in a cleavable rock lie parallel to the elongation of the rock mass as developed during rock flowage. This relationship has been so generally assumed by geologists that at first thought it would seem entirely superfluous to present evidence in proof of it. But it has been questioned by able geologists. Becker [2] has held that the elongation of the rock mass may be inclined to the common direction of the major axes of the mineral particles. The student, when asked how he knows that cleavage is parallel to rock elongation, is often completely at sea. It is simply a matter of observation. The evidence is as follows: (*A*) *Distortion of pebbles of a conglomerate.* Schistose conglomerates show by the distortion of their pebbles the plane of elongation, although it may sometimes be difficult to distinguish the shapes of undeformed pebbles from the deformed ones. The cleavage of the matrix

[1] Becker, G. F., and Day, A. L., The linear force of growing crystals: Proc. Wash. Acad. Sci., vol. 7, 1905, pp. 283–288.

[2] Becker, G. F., Current theories of slaty cleavage: Am. Jour. Sci., 4th ser., vol. 24, 1907, pp. 7–10.

is approximately parallel to the greater diameters of the flattened pebbles, although it curves somewhat at the ends of the pebbles. (*B*) *Distortion of mineral crystals.* The plane of cleavage is marked by mica plates or hornblende crystals, while the associated quartz and feldspar particles may be fractured obliquely to the plane of cleavage. The displacement of the parts, which often accompanies such fractures, is observed to extend the fractured parts in the plane of rock cleavage. (*C*) *Distortion of volcanic textures.* The original ellipsoidal partings of basalts frequently show a flattening, with or without fracture; in such cases the ellipsoids and the matrix have a flow cleavage parallel to the longer diameters. The elongation of amygdules and spherulites in planes parallel to the rock cleavage is likewise of common occurrence. (*D*) *Distortion of fossils.* The elongation of fossils in the plane of cleavage has been observed in cleavable rocks. (*E*) *Distortion of beds and attitude of folds.* Folds often show the direction of elongation of the deformed rock mass, and cleavage is commonly parallel to their axial planes (see Chapter VIII). (*F*) *Relations to intrusives.* Intrusions of great masses of igneous rocks, and particularly deep-seated batholiths, exert pressure against their walls. Any cleavage developed in the surrounding rocks is parallel to the periphery of the intrusive masses.

It is concluded then that the longer dimensions of mineral constituents are parallel to the directions or planes of elongation of the rock mass. Thus an adequate statement of the relations of rock cleavage to the stresses which have produced it must be a statement which will cover the various ways in which stress has elongated and shortened rock masses.

(2) Knowing then the relations of cleavage to elongation, it becomes possible to state its relations to stress. In the simplest possible terms (see pp. 18–19), stress has been effective in distorting rock masses (*A*) by non-rotational strain, in which the axes of stress and strain remain constantly parallel throughout the deformation, and (*B*) by rotational

strain, in which there is a continuous angular change in the position of the strain axes as compared with the stress axes during the distortion. In the first case the elongation of the rock mass is normal to the greatest stress and remains so through the deformation; in the second case the elongation of the rock mass is constantly changing in direction with reference to the principal stress, and ultimately the elongation may be considerably inclined to the maximum stress. It is held by Hoskins [1] that at any instant the tendency for elongation is approximately normal to the greatest stress, but that the rotational tendency results in inclining the final elongation to the greatest stress.

Substituting rock cleavage for greatest elongation of the rock mass, the statement of the relations of cleavage to pressure is as follows: In a non-rotational strain cleavage is developed normal to the greatest stress; in rotational strain, while at any instant there may be a tendency for it to be developed normal to the greatest stress, there is here a rotational element which brings it into position inclined to the greatest stress. All compressional strains in rock masses belong to these classes, rotational and non-rotational, and usually to some combination of the two. Cleavage, therefore, is developed under some combination of rotational and non-rotational strains and may be said to be produced both normal and inclined to pressures.

Inferences from field observations as to the stress conditions related to cleavage are discussed in connection with folds.

Relation of Flow Cleavage to Bedding and to Folding

Flow cleavage is a common accompaniment of folding, and has such uniform relations to folding that it becomes of diagnostic value in field work. This subject is discussed principally in connection with folds, but in summary we may state that the existence of cleavage being, in itself, evidence of

[1] Hoskins, L. M., Flow and fracture of rocks as related to structure: 16th Ann. Rept., U. S. Geol. Survey, pt. 1, 1896, pp. 845–874.

structural weakness, is also evidence that the rock has probably folded; that flow cleavage in general is parallel to the axial planes of the folds; that its inclination to bedding shows the direction of differential movement, which, in turn, is characteristic of certain parts of folds. With differential movement known, it is possible to see what part of the fold it is on and to identify top and bottom of the beds. The trace of bedding on

FIG. 47. Slaty cleavage parallel to the axial plane of a fold. Two miles south of Walland, Tenn. After Keith.

the cleavage planes indicates the pitch of the fold. In fact, in districts where outcrops are few or where inferences have to be drawn from drill cores, an understanding of the common relations between flow cleavage and folding is essential to diagnosis of the rock structure.

If cleavage is usually parallel to the axial planes of folds, it follows that it is usually inclined to folded bedding planes

or to other folded plane structures, such as the surfaces of sills. Where the folds are closely compressed the inclination of the cleavage to the bedding on the limbs may be very slight, though even here it is likely to exist. In wide observations through the pre-Cambrian rocks the writer has never

FIG. 48. Cleavage crossing bedding of slates, St. Louis river, Minnesota. The broad, plane surface dipping to the right is a bedding plane. The structure dipping more steeply to the right is cleavage, which is parallel to the axial plane of the fold.

yet found cleavage exactly parallel to the bedding for any considerable distance. When followed out in strike or dip for even a few yards, it is found to cut the bedding. Its low angle with the bedding is explained by the large rotational element in the ever-present shear along the beds, — which brings the longer axis of the mineral particles or the strain

ellipsoid nearly but not quite parallel to the bedding. It is conceivable that in closely compressed folds, where the limbs are parallel to the axial planes, cleavage might be exactly parallel to the limbs. If this situation exists, it is exceptional and local.

Others have reported cases of flow cleavage parallel to the bedding which they ascribe, not to differential movement, during folding, parallel to the beds, but to load on the beds while they are lying flat. A conspicuous case is cited by Daly [1] in the Shuswap terrane in British Columbia, where quartzites, limestones, shales, sills, laccoliths, and larger igneous masses all have a cleavage parallel to the bedding of the sediments. Furthermore, this prevailing cleavage cuts across dikes which are normal to the bedding planes. The dips are prevailingly low and there is an absence of sharp folds of the kind made by ordinary orogenic disturbances. Daly concludes that the flow cleavage in this case is the result of "static metamorphism" under the vertical gravitational load of overlying sediments. He recognizes the fact that the thickness of overlying sediments in this case has not been greater than that which has overlain other great pre-Cambrian terranes which do not show flow cleavage parallel to the bedding; and suggests that the effect of load on the Shuswap series was locally augmented by the salic quality of the granitic injections, by the unrivaled abundance of pegmatite, by the prevalence of sill and lit par lit injection, and in general by the highly hydrous character of the intrusives.

Some shales possess a fine fissility parallel to the bedding, which is in part the result of deposition, but which may also be due to gravitational compression and to plastic flow down the depositional slope when soft.[2]

The Grenville series of the Adirondacks consists of

[1] Daly, R. A., A geological reconnaissance between Golden and Kamloops, B. C., along the Canadian Pacific Railway: Memoir 68, Dept. of Mines, Can., 1915, pp. 44–49.

[2] Lewis, J. Volney, The fissility of shale: Bull. Geol. Soc. Am., vol. 34, 1923 (abstract).

thoroughly crystallized limestones, sandstones, and shales, with a more or less well developed schistosity parallel to the stratification surfaces. Schistosity crossing the bedding is reported as not common. The dips are in general low and intense folding is not common. Furthermore, inclusions of Grenville sediments in later intrusives obviously possessed their schistosity before the intrusions, and before the deformation which accompanied and followed the intrusions. Miller [1] concludes that the Grenville series was developed during the recrystallization of essentially horizontal strata under heavy load of overlying material, where conditions of warmth and moisture were also favorable.

Flow cleavage in this and other similar cases is not highly developed; the more conspicuous structure is a gneissic banding, in which true flow cleavage is a rude one or altogether lacking.

It is very doubtful whether such a structure is caused solely by great depth of burial, for the reason that other extensive terranes which have been, so far as now known, fully as deeply buried, do not show this structure. For instance, in the Lake Superior region the most ancient rocks are known to have been overlain at one time by many miles of sediments. Excessively deep burial can be proved in this case much more certainly than in the case of either the Grenville or Shuswap terranes, and yet these old Lake Superior rocks shows no trace of *load* or static metamorphism. All of the major secondary deformation is clearly of an orogenic type.

Whatever the explanation of the particular cases above cited, in general within our zone of observation load has not produced "static" cleavage on any large scale. It is certain that rock cleavage is commonly a result of movement. It is associated with folds, joints, faults, and other structures uniformly indicating movement. The distortion of original structures and the presence of recrystallization, granulation,

[1] Miller, William J., Origin of foliation in the pre-Cambrian rocks of northern New York: Jour. Geol., vol. 24, 1916, p. 597.

and rotation in cleavage all clearly indicate movement, — which means the shortening of certain strain axes and lengthening of others. Load could clearly accomplish movement where there is opportunity for extension normal to compression, that is, where there is opportunity for escape. For instance, lean shale formations with exposed edges are known to flow under gravitational load deep below the surface is absent, and gravity is translated into movement, however, it seems doubtful whether a cleavage can develop. In general it might seem that the opportunity for the lateral extension of the sediments under gravitational load deep below the surface is absent, and that any movement caused by load is likely to consist of local readjustments along planes for the most part inclined to the normal pull of gravity.

General Distribution and Attitude of Flow Cleavage

In some districts cleavage is confined to narrow zones of weak rock and is absent in the more massive and competent formations. In others all of the rocks may be more or less cleavable and the cleavage surfaces are substantially parallel. This is called a *regional cleavage.* The term is a relative one, for there is no definite demarcation between local and regional cleavage. What one observer might call regional, another might call local, depending upon the point of view and the detail in which the work is done.

Good examples of regional cleavage are to be seen in the Lake Superior pre-Cambrian and adjacent parts of the pre-Cambrian shield of Canada, the southern part of the Piedmont plateau of the southeastern United States, and the Moine schist area of northern Scotland. The cleavage in each of these areas has more or less uniform strike and dip, even though the rock layers are much folded and have great diversity of attitude.

Even in areas of regional cleavage it is seldom that all of the rocks are highly cleavable. The softer rocks are so, but

the stronger, more competent ones are likely to be only slightly or not at all cleavable. In fact, so far as the writer's observation has gone, the zones of highly cleavable rocks are comparatively narrow, and wind their way between considerably larger areas of massive rocks in which cleavage is not a conspicuous structure. In the oldest Archean rocks of northern Minnesota, in which the folding and mashing have probably been as severe as in any place on earth, the parts which are highly cleavable or schistose probably do not constitute a third of the total area.

The stratigraphic distribution of cleavage is highly irregular. It appears both in rocks recently deformed near the surface and in rocks of much more ancient deformation which now lie near the surface because of erosion. Older formations are likely to show more cleavage than younger ones, but it is not known whether this is because they have been more deeply buried or because they have been deformed more times. When allowance is made for repeated deformations of older formations, there is no direct evidence in the zone of observation that below a thin superficial zone of fracture the process of cleavage-making increases with depth.

As cleavage is the best evidence of rock flowage, it follows from the foregoing that our zone of observation does not disclose any considerable zone of rock flowage of the kind which has been assumed to be prevalent deep below the surface. Whether or not rock flowage is more prevalent at greater depths is a question involving many other general considerations which are treated in Chapter XIV.

There is another fact of interest relating to the general distribution of cleavage. It commonly occurs in areas where there has been much deformation of a mountain-making kind. On the other hand, there are mountains in which movements have been intense where cleavage is lacking, even in some of the softer, shaly rocks. Also there are formations in which evidence of conspicuous movement is lacking, the beds remaining almost in their original horizontal positions,

which still possess a regional cleavage, in this case more or less parallel to the bedding. These areas are usually associated with plutonic intrusion, and the cleavage partakes of the nature of a gneissic structure; it is for the most part not typical flow cleavage. It has been supposed that such cleavage may be due to static metamorphism under load, but it is difficult to eliminate from consideration the effect of intrusives and the possibility of actual movement.

In general, flow cleavage may be regarded as an exceedingly local phenomenon, occurring in narrow zones, mainly in weak formations, and anastomosing, winding, and branching between zones of more massive rocks. It represents merely local zones of weakness along which the conditions have been favorable for rock flowage. It does not represent all the zones of weakness, because some are marked by fracture, and others were unconsolidated at the time of movement and did not develop cleavage, and still others, like marble, flowed without causing cleavage. Most of it is related clearly to dynamic movement, some of it to intrusion, and some of it doubtfully to load. Where cleavage occurs around the periphery of a batholithic intrusion it is sometimes difficult to be sure whether the pressure should be classed as dynamic or that of intrusion, or both.

The strike and dip of regional cleavage is known to be rudely parallel with that of the axial planes of the folds of mountain chains with which it is commonly associated — not only existing mountain chains, but old ones which have been eroded and are now represented only by their roots. For instance, so far as there is original cleavage in the Appalachian Mountains it tends to strike parallel to the range and to dip eastward, parallel to the axial planes of associated folds. In the pre-Cambrian shield of Canada and adjacent parts of the United States there are roots of mountain chains which have entirely disappeared. These in general have a strike somewhat more easterly than the Appalachians and seem to have joined the Appalachian chain tangentially in northeastern

Canada. The cleavage has the same trend. A cross-section from Lake Superior to Hudson Bay shows remarkable uniformity in cleavage directions. The above generalization is tentative, for as yet observations have not been collected and correlated from all parts of the world.

Cleavage being parallel to the axial planes of the folds of mountain chains, and some of these in turn being parallel to continental margins, it follows that cleavage to that extent is also parallel to continental margins.

Willis [1] has developed the hypothesis that the unloading and loading effects of erosion are competent to orient cleavage, and that sub-oceanic cleavage is oriented in approximately horizontal attitude, but rises in the broad marginal regions to join the curve that extends up into the eroded continental areas. He calls the mass characterized by flat foliation a disc and the mass in which the foliation rises to a steeply dipping attitude an inter-disc. Whatever the merits of this reasoning, the sub-oceanic attitude of cleavage here assumed cannot be based on direct observation.

Cleavage which is clearly the result of dynamic action, as distinguished from gneissic and load cleavage, has prevailingly high dips, steeper than the bedding. This generalization applies to practically all of the pre-Cambrian shield of North America. Locally, in areas of great overthrust, the cleavage has lower dips, but still usually steeper than the bedding. Low-dip cleavage is a natural result of shear tangential to the earth's surface. Vertical cleavage means either non-rotational stresses applied normal to the plane of cleavage and parallel to the earth's surface, or shearing stresses inclined at any angle to the cleavage planes. In the latter case, they may be due to horizontal movements of different parts of the earth at different speeds, — of the kind, for instance, which have been noted in connection with geodetic observations in California, — which result in shearing stresses acting tangentially

[1] Willis, Bailey, Discoidal structure of the lithosphere: Bull. Geol. Soc. Am., vol. 31, 1920, pp. 247–302.

in a horizontal direction on vertical planes. In general, then, the prevailing attitudes of cleavage, so far as they are known, can all be plausibly accounted for by horizontal non-rotational compression or by horizontal shearing stresses applied at various angles to the resulting planes of cleavage.

Cleavage is much more likely to have a uniform attitude than the zone of movement of which it is a part.

To illustrate: Cleavage may be confined to a folded slate formation between two non-cleavable formations. The strike and dip of the slate horizon vary from place to place with the folding, but so far as the axial planes of the folds are more or less parallel, as they usually are in regional deformation, the cleavage is likewise more or less uniform in its strike and dip. In general, where the cleavage has a strong tendency toward verticality, the zones of movement marked by the cleavage may have almost any inclination or direction. Any regional deforming forces tangential to the earth's surface have been resolved along movement zones at various angles to the applied forces, and it is difficult to conceive of the absence of rotational movements under these circumstances.

Much of the discussion of the general attitude of cleavage is more or less influenced by *a priori* acceptance of some hypothesis of earth deformation. Observations on cleavage itself are not sufficiently widespread nor sufficiently well correlated, and disclose so many exceptions and irregularities, that as yet cleavage does not itself constitute a safe basis for strictly inductive reasoning.

Grain and False Cleavage

Slates and schists often have a direction of easy breakage across the flow cleavage, called *grain* or *false cleavage*, which is utilized in quarry operations. It may be characterized by striations of the surfaces of flow cleavage in a direction nearly parallel to the cleavage dip. In some cases grain follows the plane of the greatest and least axes of the parallel

mineral particles arranged by flow cleavage. The best breakage, that of the flow cleavage itself, follows the plane in which lie the greatest and mean axes. In other cases grain seems to be fracture cleavage or false cleavage marked by minute monoclinal faulting and folding of the flow-cleavage lamellae, in planes variously inclined to the flow cleavage (see Fig. 55).

Idiomorphic or Porphyritic Textures Accompanying Flow Cleavage

Garnet, staurolite, tourmaline, andalusite, chloritoid, and other heavy anhydrous minerals of this kind are usually idiomorphic or porphyritic in cleavable rocks. They develop by recrystallization after rock flowage has ceased, but probably while the rock is still under high pressure and temperature, as is evidenced by their high specific gravity and frequent occurrence in the proximity of intrusive igneous rocks. Their late development by recrystallization is shown by the following considerations: (1) They appear in rocks clearly derived by rock flowage from others originally lacking such minerals. (2) They frequently lie at large angles to the prevailing cleavage in the rock. (3) They do not show the degree of mechanical deformation that they would necessarily have possessed had they developed their present positions before flowage had ceased. Many of the crystals are long and acicular, and would surely have been broken if any considerable movement had occurred subsequent to their development. (4) They include, within their boundaries, minerals in part similar to those in the remainder of the rock, and which have an arrangement of their greater diameters in the plane of rock cleavage, — showing that to some degree at least such minerals were formed during rock flowage and that the porphyritic developments came later. (5) The mica and the other constituents of cleavable rocks, which are certainly developed by recrystallization during the process of deformation, are frequently seen to end abruptly at the periphery of

a mineral of this group and not to curve around it as they often do about the resistant minerals in schists. If the rock had flowed after the formation of the porphyritic crystals, crowding and bending of the micas must inevitably have occurred. (6) The large size of minerals of this group, as compared with their associated mineral particles, suggests

Fig. 49. Photomicrograph of chloritoid crystal in micaceous and quartzose schist from Black Hills. The chloritoid crystal here shown has developed later than the rock flowage producing the prevailing cleavage of the rock. The chloritoid has grown at the expense of the other constituents of the rock, using all the material necessary for its growth and leaving the excess of material in the form of inclusions, which retain their dimensional parallelism with the prevailing rock cleavage. Bull. 239, U. S. Geol. Survey.

their development subsequent to rock flowage, when granulation is no longer tending to break down the crystals.

For these reasons the development of this group of crystals is believed to have been mainly later than the formation of

the cleavage, but it is true also that in some cases subsequent flowage has fractured them and bent the other constituents around them. The very fact that the effects of further movement, when it occurs, are so conspicuous, confirms the conclusion that the secondary porphyritic minerals not showing these effects developed after the movement ceased.

We may only speculate as to the conditions of this peculiar development of non-arranged minerals. They may be those of high temperature and pressure with uniform, or cubic compression (p. 18) but apparently not of movement. Possibly the pressure and temperature may be considered as having become so great as to develop hydrostatic conditions, in which case there would be none of the differential pressures necessary for a parallel arrangement of constituents.

Another possible explanation is suggested by the experimental fact that rocks under stress, but with lateral support, take on a great rigidity, which makes it possible for them to resist considerable stress differences without yielding. It seems possible that the schists, at the time of development of the porphyritic constituents, may have been under great differential stresses; but because of the high rigidity, due to lateral support, may have been able to withstand these stresses and not yield further by rock flowage. It may be supposed that the progress of rock flowage tended to lower the stress difference to a point where it was not great enough, with the concomitant increase of rigidity, to cause further movement. These conditions would differ from hydrostatic conditions in that there would still be considerable differential stress, but not sufficient to cause movement.

Whatever the explanation, the uniformly high density of the porphyritic constituents suggests that they have formed in response to demand for less volume.

Obliteration of Primary Textures by Flow Cleavage

Recrystallization, the dominant process in rock flowage which causes cleavage, tends toward an increase in the size of grain, the segregation of minerals into bands, a uniformity in size and shape of the mineral particles, and the growth of new minerals such as mica or hornblende not previously existent in the rock. Previous textures are commonly destroyed. Bedding is locally not completely obliterated, because alternation of beds of originally different mineralogic composition and texture determines to some extent the kinds and sizes of the secondary mineral particles formed in these beds by rock flowage. Thus a faint banding of dark or light minerals, or of fine or coarse minerals, may mark the original bedding in a schistose rock. The perfection of cleavage is determined largely by the original nature of the bed, with the result that in following cleavage from one bed to another there are likely to be marked differences in its texture. In exploration for commercial slates it is necessary to follow the horizons of originally homogeneous beds rather than to follow the cleavage itself along its strike and dip.

FIG. 50. Cleavage crossing bedding in slate. The vertical lines are cleavage; the dark, inclined bands are bedding.

In exceptional cases where the cleavage is parallel to bedding, noted elsewhere, the primary structures may be retained or emphasized.

Flow Cleavage as a "Crystalloblastic Structure"

Rocks with flow cleavage have a *crystalloblastic* structure. In the foregoing description we have avoided the use of this

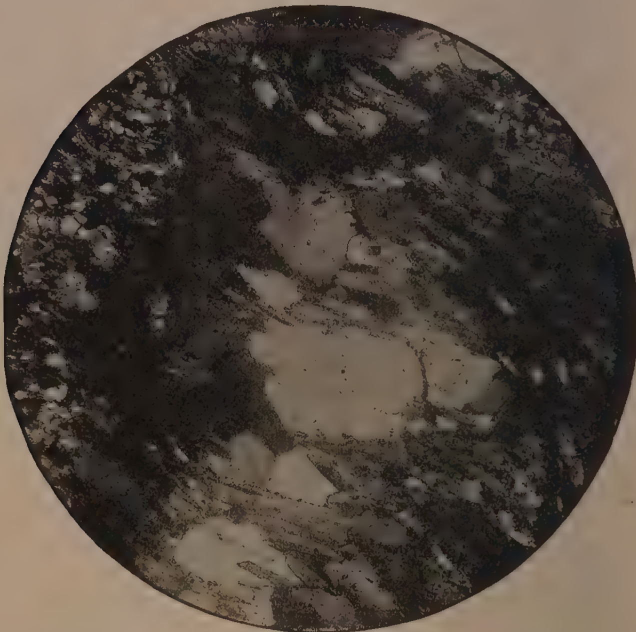

FIG. 51. Photomicrograph of micaceous and quartzose schist with cleavage developed across original bedding, from Little Falls, Minn. A graywacke-slate, in which the banding has been marked by difference in texture as well as in composition, has been subjected to deformation, with the result that a cleavage has been superposed at right angles to the original bedding. Originally the longer diameters of the particles of the bedded rock were parallel to the bedding. Accompanying the development of flow cleavage most of the constituents of the rock have been recrystallized. The quartz particles shown in the light band have been drawn out with their longer diameters nearly at right angles to the former plane of their longer diameters, and abundant new mica has developed with its greater diameters and mineral cleavage normal to the plane of bedding. Bull. 239, U. S. Geol. Survey.

term because its definition includes certain considerations to which we do not entirely agree.

Becke [1] distinguishes between characteristics of textures of igneous rocks and those of schists. The texture of igneous rock is controlled by successive crystallization of minerals. In crystalline schist there is no successive crystallization. There is developed the so-called crystalloblastic texture, the characteristics of which are as follows: (1) The constituents are of equal rank and show no definite order of crystallization. Any one mineral may occasionally exist as an inclusion within another. The later-formed minerals may have more perfect crystal outlines than their inclusions. (2) Crystal forms are rare. Those present are usually simple, and there is marked development parallel to the cleavage planes. (3) Lack of skeleton crystals. (4) The minerals in the schists with the most compact molecular arrangement are most likely to take on crystal form, that is, have the strongest crystallizing power; and on this basis Becke develops a series of "crystalloblastic" minerals named in order of their decreasing crystallizing power. The usual series of form development is as follows: titanite, rutile, hematite, ilmenite, garnet, tourmaline, staurolite, cyanite-epidote, zoisite-pyroxene, hornblende-magnesite, dolomite, albite, mica, chlorite, talc-calcite-quartz, plagioclase-orthoclase, microcline. (5) Parallel structure of the crystalline schists is the result of rotation of already existing minerals and of recrystallization. (6) Crystalloblastic structure is holocrystalline, not cellular. (7) Zonal structure in minerals is absent, or, if present, follows different rules than in igneous rocks. (8) Inclusions do not follow the strata of the crystal but are related to the pyramids added to the original crystal or else to an older helizitic structure.

From our standpoint this statement of the principal features of secondary schistose and gneissic textures is essentially correct, but we would suggest certain qualifications, principally in regard to the statement that the minerals in schists

[1] Becke, F., Über Mineralbestand und Struktur der kristallinischen Schiefer: Compt. Rend., IX Cong. Geol. Internat., Vienna, 1903, pp. 553 *et seq.*

develop in a crystalloblastic order determined by their crystallizing power and density. In general, it may be agreed that minerals in schists with the best crystal development have the strongest crystallizing power and are in general dense, but that this is due to their crystallizing power or to their density alone does not necessarily follow. As indicated on earlier pages, during the movement of the rock mass the development of new minerals seems to be influenced by crystal habit or dimensions, — for during this phase of the development of schists and gneisses only columnar and platy minerals are developed, adapted dimensionally to the conditions of movement required by the unequal pressures. Later there may be porphyritic development of garnet, tourmaline, staurolite, and other heavy, anhydrous minerals, arranged entirely independent of the schistosity, indicating that there has been no movement during or subsequent to their development, and that they have probably developed under mass-static conditions. The high average density of these minerals suggests that density may have been a factor in their development.

The statement that the mineral constituents of schists are of equal rank and show no definite order of crystallization may also be questioned. The order seems to be a fairly definite one: first, the development of platy and columnar minerals during rock movement; and later, the development of porphyritic anhydrous constituents, where there is no movement. Each mineral newly developing in a schist derives its material from the destruction of preëxisting minerals. This differs from the conditions in igneous rocks, where minerals crystallize from a magma, and those with the strongest crystallizing power presumably crystallize first, the remainder filling the spaces between.

Finally the statement may be questioned that crystal forms are rare. The characteristic development of platy, columnar, and porphyritic minerals in the schists, with their own distinctive habits, means a uniform approach to crystal form, —

which does not seem to the writer to be less marked than in igneous rocks, where many of the later developments fill the spaces between the early developed crystals, and, therefore, cannot take on their own crystal form. So far as the writer's observations and measurements go, the tendency is for the common minerals of the schists to take on crystal habits not far different from those in the igneous rocks.[1] Trueman [2] also noted a similarity in habit.

General Comments on the Mineral and Chemical Changes Involved in the Production of Rock Flowage

Igneous rocks under rock flowage may pass into schists and gneisses similar in all essential respects to those formed from the sediments. Whether the parent rock be igneous or sedimentary or whatever its mineral content, the resulting schists and gneisses are characterized by hornblende, chlorite, and mica, in some cases, feldspar, which are developed to such an extent that the very nature of the original rock is often lost. If the mineralogical changes were not extensive, the problem of origin of schists and gneisses would not be nearly so difficult as it is. The very existence of the problem testifies to the great mineralogical changes which have occurred.

It is inferred from the available facts [3] that, while recrystallization of substances present has of course played an important part in the production of schists and gneisses, for some rocks important changes in composition have also occurred; that these changes have tended to give the rock a composition in which the influence of mica, hornblende, or chlorite is clearly discernible, that these changes are known both in sedimentary and igneous rocks, of both acid and basic composition, and

[1] See Leith, C. K., Rock cleavage: Bull. 239, U. S. Geol. Survey, 1905, pp. 24–48.

[2] Trueman, J. D., The value of certain criteria for the determination of the origin of foliated crystalline rocks: Jour. Geol., vol. 20, 1912, pp. 236–241.

[3] Leith, C. K., and Mead, W. J., Metamorphic geology: Henry Holt & Co., New York, 1915, p. 201.

that the changes have been sufficiently important to make it impossible, along with other reasons, to use chemical composition as a conclusive criterion for the identification of origin of schists and gneisses. Obviously these minerals are adapted to the conditions of rock flowage; otherwise, they would not develop at the expense of other minerals. It is not clear whether they are adapted by their crystal habits, by their cleavage, by their composition, by the temperatures required for their development, or by a combination of these qualities.

Rock Flowage Without Cleavage

Marble is the commonest example of a hard rock which undergoes flowage without retaining cleavage. It often occurs between schistose beds which have flowed; its bedding is contorted; without doubt the marble itself has flowed, and yet it possesses no cleavage. Cleavage may be produced experimentally in marble by pressure alone, when the conditions are not favorable for recrystallization.[1] Microscopic examination indicates that this has been accomplished by granulation, slicing, and gliding of the calcite crystals. Occasionally such a cleavage is observed in marble deformed under natural conditions. It may be supposed that many marbles have shown this structure in early stages of their flowage, but calcite recrystallizes so easily that the parallel structure caused by mechanical deformation is soon destroyed. The recrystallized calcite crystals do not have the habit necessary for a good dimensional arrangement in schists.

So far as limestones or marbles have impurities in them, secondary silicates are likely to develop, such as actinolite and tremolite, which by their arrangement may give the rock a cleavage.

In contact zones about large plutonic igneous masses, rocks

[1] Adams, F. D., and Nicolson, J. T., An experimental investigation into the flow of marble: Phil. Trans. Roy. Soc. London, vol. 195, 1901, pp. 363–401. See also: Adams, F. D., and Coker, E. G., The flow of marble: Am. Jour. Sci., vol. 29, 1910, pp. 465–487.

may be thoroughly recrystallized during rock flowage into masses without cleavage, though sometimes with a rude banding, — hornstones, gneisses, and similar rocks being thus produced even from such refractory substances as quartzites and shales.

The flowage of unconsolidated formations like mud, or sand, or marl, does not produce cleavage, though there is much distortion expressed in folds and faults. After such a distorted rock is hardened it is not easy to determine whether the flowage took place in the soft or hard rock stage, but the absence of cleavage may suggest that the flowage occurred before hardening. Criteria are not yet decisive (see pp. 228–231). For all we know, there may be folding in the hard rock stage, accomplished by minor granulation and recrystallization, without the development of cleavage.

Gneissic Structure

Gneissic structure consists of a rather coarse banding of feldspar and quartz with mica, chlorite, hornblende and other minerals, with or without the parallel dimensional arrangement necessary for rock cleavage or schistosity. A gneiss differs from a schist or slate in the coarseness of its banding and grain, in the usual abundance of feldspar, and in the fact that true flow cleavage or schistosity dependent on the parallel arrangement of columnar or platy minerals is not an essential characteristic.

While a schistose or slaty structure is always the result of structural deformation, the gneissic structure is known to originate both in this and other ways. So far as gneisses are the result of secondary deformation, they show in some degree the characteristics of flow cleavage already described. Gneisses formed in other ways may also have some structural characteristics of flow cleavage, but in varying degrees, and in addition they have other characteristics, due to their difference of origin.

The problem of the origin of gneisses in its relation to deformation is a large and intricate one, involving consideration of composition, magmatic conditions, metamorphism, and other conditions, as well as those of structure. We shall here confine our attention to a few elementary considerations touching most closely the structural field.

The origin of many gneisses is not yet known, but among the gneisses the following principal types have been recognized:

(1) Some gneisses have been identified as resulting from flow while still in the molten stage. These gneisses are called *protogene gneisses* or *primary gneisses*. In general these differ from schists in their higher content of quartz and feldspar, in coarseness of banding, and in the lesser tendency to a parallel arrangement of the mineral particles. Cleavage parallel to the banding is usually poor. This structure is found in many batholithic areas under conditions which show that it is not the result of pressure acting on the rock after it is cold. The banding may be parallel to the periphery of the batholith, following even detailed irregularities of the wall rock and at an angle to the schistosity of the enclosing rock. Associated pegmatitic dikes, which are clearly after-effects of the same intrusion, may cut across the gneissic structure, showing the very early period of development of the gneissic structure. These gneissic batholiths sometimes include foreign blocks which either have not been deformed at all or have an earlier schistosity which lies at an angle to the main gneissic structure of the enclosing rock. The banding does not follow a plane surface for any considerable distance, but branches, curves, and is otherwise irregular, suggestive of magmatic flowage. The minerals and textures are of a kind characteristic of crystallization from igneous melts. Some dikes show a parallelism of banding to wall. (See also pp. 231–232).

The cause of primary gneissic structure of this kind is clearly movement of the magma before it solidifies, arranging the minerals parallel to the magmatic currents; but in some cases there is a gradation into structures formed later, and it is

not easy to be certain of the origin. There are cases where primary banding is rudely parallel to the surfaces of batholithic domes, but where for a few feet or yards at the contact the banding passes into a finer schistose structure which may be parallel to the schistosity of the surrounding rocks, — indicating a pressure effect which may have been exerted in part later than the primary banding, though probably still in the later stages of the main upward thrust of the magma mass.

(2) Bedded impure quartzites and shales are known to recrystallize into gneisses under extreme conditions of anamorphism, particularly where invaded by abundant igneous intrusives. The banding is due to original differences in composition of the beds. Flow cleavage is not conspicuous, and so far as it is present, it is as likely as not to be at an angle to the bedding. In general, the banding of such gneisses is much more uniform than that of the primary gneisses. Sometimes the mineralogical and chemical composition of the rock may disclose sedimentary origin, though not always. A clue to the origin of such gneisses may be afforded by their common interbanding with limestone, quartzite, or graphitic layers, and by their transition into sediments along strike or dip.

(3) Some gneisses are formed by the penetration of magmas parallel to sedimentary banding or to other fissile structures. These are called *injection gneisses*. The process usually results in the coarse recrystallization of the original layers and their "freezing" to the intrusive material, — yielding a rock of complex qualities, which it is difficult to identify on the basis of mineralogical or chemical composition. In the field, however, it is often possible to follow one of these minutely banded rocks into coarser phases where the intruded and intrusive portions can be identified and their relations ascertained.

(4) Some gneisses result from deformation of igneous rocks, as do schists and slates. Microscopic examination shows that in some of these gneisses the process has been mainly one of slicing and granulation of particles, strewing

them out into bands without any large amount of recrystallization or development of new minerals. The sheared anorthosites or gneisses near Montreal[1] are probably an illustration of this, although the possibility is not precluded that the granulation occurred in a late molten stage. In other cases the dominant process has been recrystallization, resulting in a general coarsening of grain, in increase of platy and columnar minerals, like mica and hornblende, and in the later development of porphyritic crystals, not oriented. In general the rock exhibits all the conditions observed in a schist or slate, but in accentuated form. It is often observed that a rock which might be called a schist, when traced into conditions where metamorphism has been more intense, becomes coarser-grained, more obviously banded, and takes on a structure which might more appropriately be called gneissic.

The deformation of a granite by this process yields a gneiss, and it has been customary to emphasize this as a very common origin of gneiss. While instances have been found of the transition from granite to gneiss clearly due to secondary deformation, it is difficult to prove this origin satisfactorily on any large scale, or to collect specimens ranging from the unaltered granite to the gneiss, for the purpose of finding out exactly what changes took place during the deformation. It is easy to find places in the field where a gneiss is contiguous to a granite, but the problem is often complicated by the difficulty of eliminating other origins for the gneiss and proving that it has actually developed secondarily from the granite.

Fracture Cleavage

Fracture cleavage is a capacity to part along closely spaced parallel surfaces of fracture or near-fracture, commonly in a single set, but occasionally in intersecting sets. It is closely related to joint structure, but the joints are so closely spaced

[1] Adams, F. D., Report on the geology of a portion of the Laurentian area lying to the north of the island of Montreal: Ann. Rept., Geol. Survey of Canada, vol. 8, pt. J, 1896, p. 85 *et seq.*

as to give the rock a distinctive structure not ordinarily to be described in terms of joints. Where the fracture cleavage is in a single direction the rock is rendered minutely fissile parallel to a single surface. When in intersecting sets, the rock breaks in polygonal blocks or in parallelopipeds. Fracture cleavage differs from ordinary flow cleavage or schistosity in that the surfaces of breaking are not determined by a parallel arrangement of mineral particles, but are independent of any

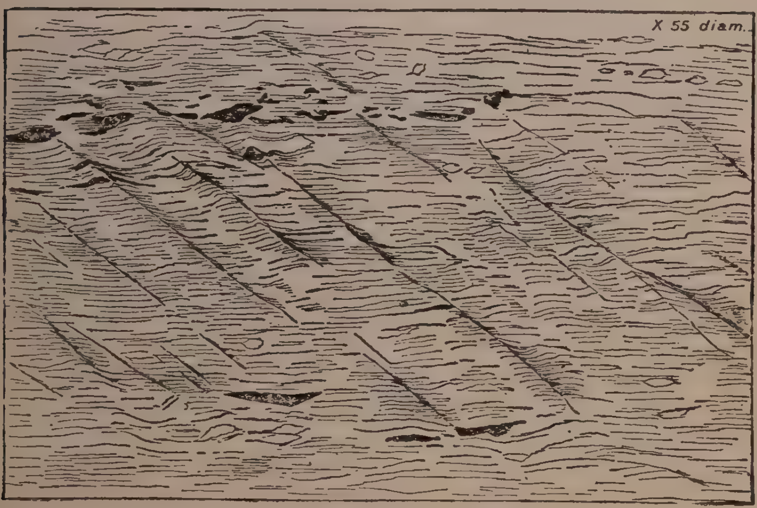

FIG. 52. Fracture cleavage crossing flow cleavage. After Dale.

such arrangement. Also it does not pervade the entire mass and affect all particles, as does flow cleavage.

Various other names have been applied to fracture cleavage, such as *fissility*, *close-joints cleavage*, *false cleavage*, *fault-slip cleavage*, *rift*, etc. These terms are not all strictly synonymous but each designates some aspect of fracture cleavage.

Fracture cleavage is often intimately related to flow cleavage in its field occurrence and origin. They may exist side by side and one structure may grade into the other. It is often difficult to determine without microscopical examination

whether the structure should be called fracture cleavage or flow cleavage. The surfaces of parting of a true flow cleavage in a softer layer may be continuous with the surfaces of parting of true fracture cleavage in a harder, more brittle layer. The flow cleavage itself, while dominantly determined by the parallel arrangement of the mineral particles, is often subordinately caused by actual shearing fractures of a fracture-cleavage type. Movements starting along fracture-cleavage

FIG. 53. Fracture cleavage developing polygonal blocks in slate previously possessing flow cleavage. Bull. 239, U. S. Geol. Survey.

planes result in the secondary growth of platy minerals along the fracture surfaces, and the result may be practically a flow cleavage. However, the existence of these gradational phenomena and the close field relation of fracture cleavage and flow cleavage should not obscure the fact that in their typical development the two structures are fundamentally different.

The fracture-cleavage surfaces may be actual open breaks separating the rock mass. They may be breaks which have been more or less cemented. They may be merely incipient or potential surfaces of fracture along which stresses have accumulated which localize fracture under further shock. These are often invisible to the naked eye. Under the mi-

croscope they may be seen to be extremely tight joints with a hair-like appearance in cross-section. Not uncommonly there has been a local distortion of minerals along these planes, indicating displacements of a fault or monocline type. This kind of structure has often been described as *false cleavage.* It is especially well developed where a fracture cleavage crosses a preëxisting flow cleavage and the parallel arrangement of minerals inherent in flow cleavage is minutely disturbed along planes of incipient fracture which cross the longer dimensions

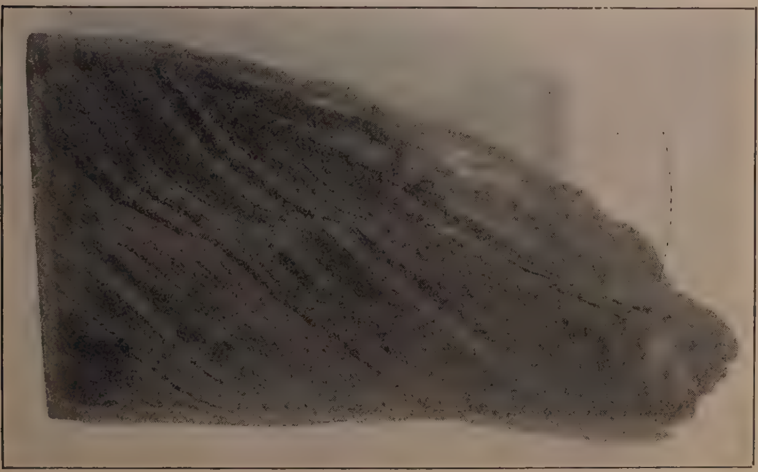

FIG. 54. Detail of a plane of fracture cleavage. There is absolute fracture in some beds and mere flow in others. Secondary mica is developed parallel to the fractures. The dark bands are bedding. After Rettger. Baraboo quartzite.

of the minerals at an angle. Fig. 55 represents the microscopical appearance of such a rock. It is not unusual to see secondary minerals, such as chlorite, mica, magnesite, pyrite, and staurolite, lying with their longer dimensions in the plane of the false cleavage. These seem to be porphyritic developments later than the flow-cleavage-making minerals.

Excessive development of false cleavage superposed on a flow cleavage may yield a minute crinkling or *shear zone,* a

name used in slate quarries. When struck with a hammer the rock will break partly along the crinkled flow-cleavage surfaces and partly following the axial planes of the crinkles.

In general when a fracture cleavage is superposed on a flow cleavage it modifies the structure but does not destroy it. On the other hand, when a flow cleavage is superposed on a frac-

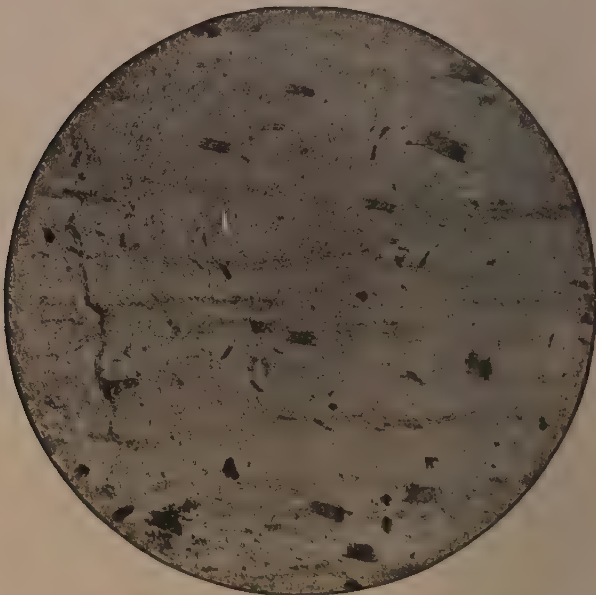

FIG. 55. Photomicrograph of slate with false or fracture cleavage from Black Hills of South Dakota. The longer diameters of the particles, mainly mica, quartz, and feldspar, lie, for the most part, in planes intersecting the plane of the page and parallel to its longer sides; but in well-separated planes at right angles to this direction the longer diameters of the particles have been deflected into minute monoclinal folds represented by the darker cross-lines. In these cross-planes, also, porphyritic biotites have developed with their longer diameters parallel. The rock has two cleavages, one conditioned by the prevailing dimensional arrangement of the minute particles and the other conditioned by the planes of weakness along the axes of the minute monoclinal folds crossing the prevailing cleavage. The first cleavage is flow cleavage developed in normal fashion during rock flowage, and the second is of the nature of fracture cleavage developed later along separated shearing planes. The rock cleaves into parallelopiped blocks. Bull. 239, U. S. Geol. Survey.

ture cleavage the earlier structure is usually completely destroyed.

The most typical development of fracture cleavage may be seen where beds of differing composition and competency, as, for instance, quartzite and slate, have been deformed together. This is illustrated widely in the Lake Superior region and in the Belt terranes of Idaho and Montana. In a softer, shaly layer a fracture cleavage or a true flow cleavage develops. As a parting surface is followed toward an adjoining harder

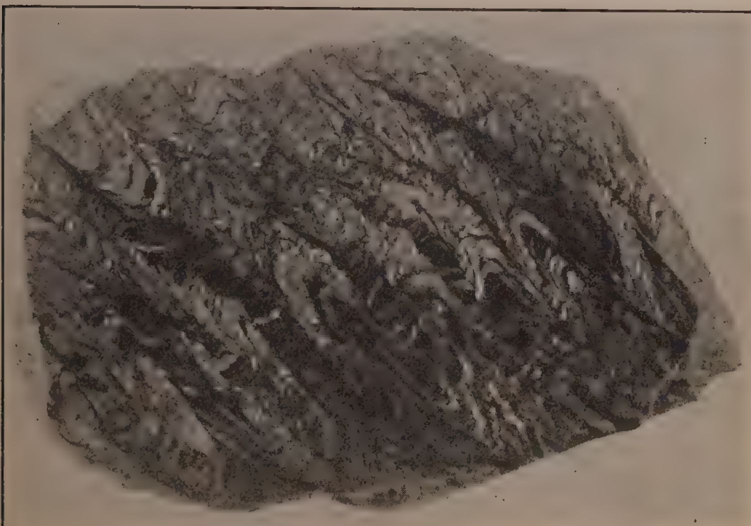

FIG. 56. Fault-slip cleavage in gneiss from southern Appalachians. The gneiss has been closely crenulated, and the minute folds may be observed to pass into minute faults which now represent planes of fracture cleavage. The faults may have been cemented or may have been welded by actual pressure. Parallel to the faults there has also been developed a parallel arrangement of the mineral particles, perhaps due in part to the slipping along the fault planes, and it is exceedingly difficult to distinguish between the fracture cleavage and the flow cleavage. Bull. 239, U. S. Geol. Survey.

layer, it is found to curve in a direction which will carry it more directly across the harder layer. In the harder layer the planes of parting are fewer and are distinctly of the

character of joints, even though the rubbing of the surfaces may have resulted in the growth of secondary micas or other minerals parallel to the surface. The partings are actually continuous with those of cleavage in the adjacent slaty layer, and are obviously caused by the same stresses. It merely means that the harder, more brittle layer is broken by joints

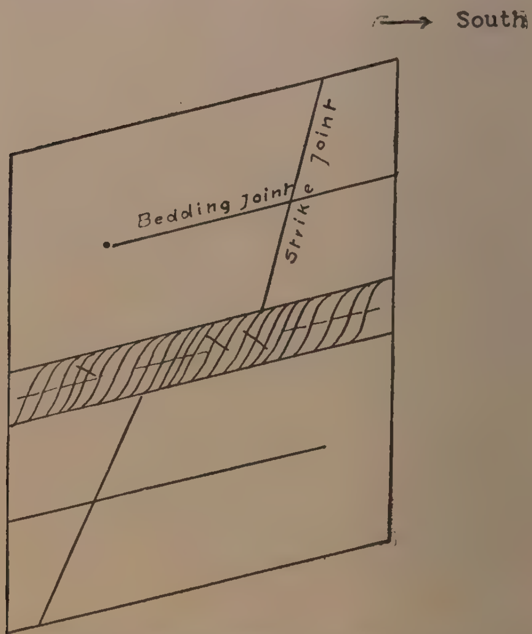

FIG. 57. Vertical section Baraboo quartzite, normal to the strike, on the South Range, Baraboo district, Wisconsin, showing fracture cleavage in shaly beds and joints in massive quartzite beds above and below. Short, open gashes or tension joints may be seen crossing the curved compression joints in the softer layers. After Steidtmann.

while the softer layer has yielded by the more pervasive structure of fracture cleavage or flow cleavage.

Fracture cleavage, in such cases, is clearly developed under rotational stress, or shear, and tends to follow a plane of no distortion; but as the rock is somewhat soft and incompetent the actual break makes a more obtuse angle with the principal

stress. The flow cleavage is developed by the same stress, but is parallel to the two longer axes of the strain ellipsoid. Figure 57 expresses these relations diagrammatically. The same relationships are expressed also in figures 58 and 59.

The angle between the flow cleavage and fracture cleavage may be so slight that it fails to serve by itself as a criterion

FIG. 58. Fracture cleavage developed in slaty quartzite layer between two massive beds of quartzite, on south limb of the Baraboo syncline, Wisconsin. The joints in the massive quartzite are continuous with the fracture-cleavage partings in the slaty quartzite. (See also Figs. 57 and 59.)

for separating the two structures. The fracture cleavage, while at an angle to flow cleavage, has qualitatively the same angular relations to bedding as flow cleavage, and like the flow cleavage it may be used as evidence of differential movement between beds, — which in turn may lead to other valuable inferences (see pp. 181–185). Where well developed, further-

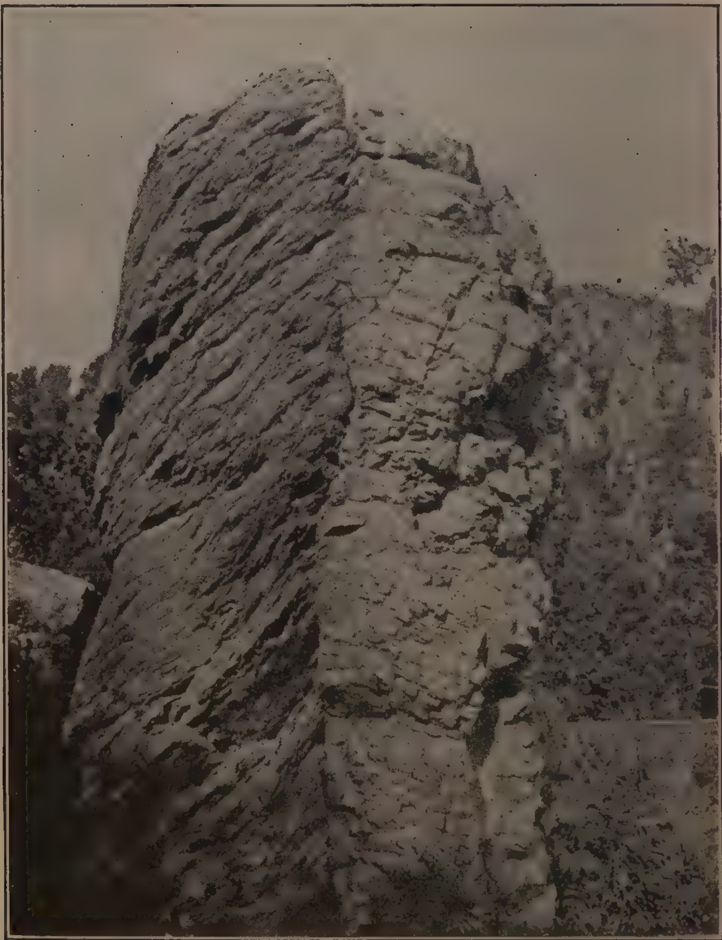

Fig. 59. Fracture cleavage and jointing developed by shearing between beds in Baraboo quartzite. "Van Hise Rock." The light portion on the right is the bed of brittle quartzite. The dark portion on the left is a bed of softer shaly quartzite. The outcrop is a part of the north limb of a syncline. The righthand bed is on the south. It has obviously moved upward with reference to the beds to the north of it, as would be expected from this position on the syncline. The fractures have been developed by rotational or shearing stresses described on pp. 18–19. It is suggested that the student superpose on these beds the theoretical positions of the strain ellipsoids and the planes of maximum shear. Note relations of fracture cleavage to jointing in adjacent bed.

more, its attitude is so nearly parallel to that of flow cleavage that it indicates roughly the position of the axial plane of the fold. Flow cleavage, as will be seen later, more exactly marks this plane. For field diagnosis, however, with qualifications,

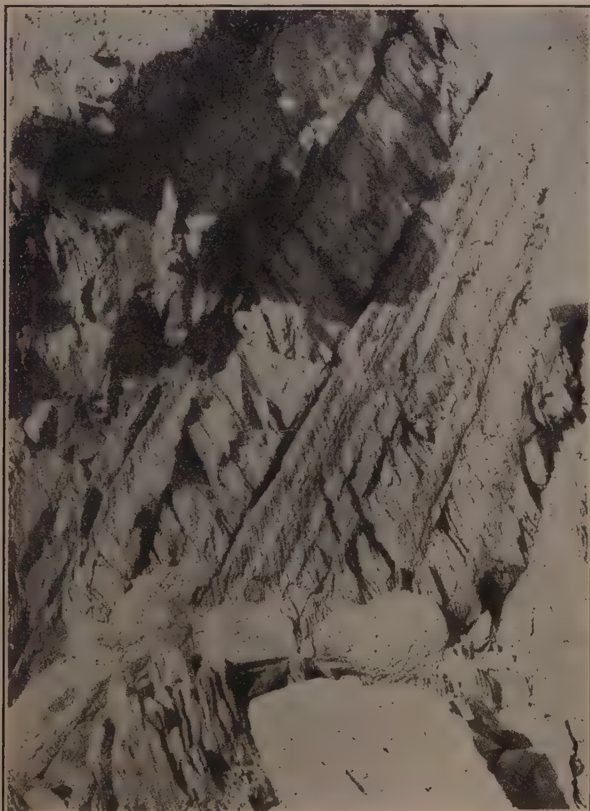

FIG. 60. Fracture cleavage, jointing, and flow cleavage developed in graywacke and slate, Alaska. After Gilbert (photograph by U. S. Geol. Survey). Use principle of strain ellipsoid (see pp. 21–27) to ascertain direction of relative displacement and theoretic position of fracture planes and flow planes.

fracture cleavage often serves much the same purpose as flow cleavage.

In a rotational shear of the sort just described there are always two intersecting shearing planes, corresponding to the two planes of non-distortion in a strain ellipsoid. One of these planes is likely to be parallel to the bedding, and any breaking along it may escape detection. The conspicuous structure always follows the other plane of no distortion, which is inclined to the bedding. The reason for the dominant expression of this shearing plane has not yet been ascertained. It probably has something to do with the differing friction and load on the two planes. Whatever the reason, while the two complementary sets of shearing stresses are always present, the rock finds it easier to yield along the one inclined to the bedding than it does along the one parallel to it.

Fracture cleavage, in the typical case described, may furnish a clue to the origin of at least part of the joints in the brittle bed; and with this as a starting point other joints, earlier or later or in other attitudes, may be ascribed to stresses other than those causing the set related to the fracture cleavage. As the fracture cleavage in the common case described is related to the differential movement between beds during folding, it becomes possible then to assign other joints to pre-folding or post-folding periods.

From the above description it is clear that the consideration of fracture cleavage cannot be separated from that of flow cleavage and of joints. In its most typical development, however, it is usually closely associated with flow cleavage.

CHAPTER VIII

FOLDS

General Description

Curvatures, flexures, and contortions of rock layers, caused by diastrophism in both igneous and sedimentary rocks, are commonly termed folds. Folds range from miles in extent to minute contortions and have all degrees of curvature. Nearly all of the sedimentary rocks of the earth possess folds, although over large areas the curvature is only slight.

Folds may form simultaneously with rock fracture or rock flowage (flow cleavage), or both, and are merely another expression of the deformation by fracture and flowage. Not all folding requires fracture or flowage, nor does all fracture or flowage involve folding. Very gentle folding may require little fracture or flowage. Even a substance as brittle as glass, in sheets commensurate in thinness and extent with many sedimentary beds, may be flexed slightly without breaking. Some folds in sedimentary formations are known to exist in a state of strain within the elastic limit, because when pressure is relieved by excavations they may suddenly change their form; they tend to spring back to their original form. Commonly, however, recrystallization gives folds a permanent set, as might be expected in view of the vast time available for this process. Folding in soft sediments or in molten masses does not require either fracturing or the kind of flowage which is registered as flow cleavage.

Certain discussions of folding in mountain ranges have implied that folding is a usual accompaniment of rock flowage rather than fracture, and, on the assumption that rock flowage increases with depth, that folding takes the place of fracture with depth. If folding results both from flowage and fracture,

however, there does not seem to be any sound basis for this inference.

Folds are formed both in soft and hard rocks. It is not easy to be sure whether a fold now observed in a hard rock was developed after the rock was hard or was developed when in an earlier soft condition (see Chapter IX).

Recrystallization, solution, and slumping all cause folds. Increase in volume in the crystallization of gypsum is the cause of some very remarkable folding. In gypsum and salt deposits in Harvey County, Kansas, there are folds 20 feet or more across, presumably to be attributed to volume changes due to these causes.

Coral reefs are nearly always associated with an up-bending of strata above and a down-bending below, — a condition presumably brought about by original deposition, but perhaps emphasized by volume changes during subsequent recrystallization and slumping.

Folds similar in all essential respects to those caused by deformation may be formed where sediments are deposited on an irregular surface, or where igneous flows adapt themselves to a curved surface. It has been found, for instance, in an exploration for oil in slightly flexed sediments of the Mississippi Valley, that some of the anticlines and domes are the result of primary deposition over anticlines or domes on the bottom, and that these structures have been later accentuated by the slump of the sediments during their drying and recrystallization.[1] While the term folding is ordinarily restricted to the structures resulting from secondary deformation, it is not always possible to be sure of this origin, and so in practice the term may include the curved structures of original deposition.

Names of folds are ordinarily based on form, but qualifying terms are often desirable to indicate the origin of the folds, the kinds of movements involved, and the relations of

[1] Powers, Sidney, Reflected buried hills and their importance in petroleum geology: Econ. Geol., vol. 17, 1922, pp. 233–259.

the folds to fracture and cleavage, which are usually other expressions of the same deforming force.

Elements of Folds

For the purpose of precise analysis and mapping of folds certain names have been applied to their parts. *Strike* is the direction of the line of intersection of a bed with a horizontal plane. *Dip* is the angle between the bed and the horizontal,

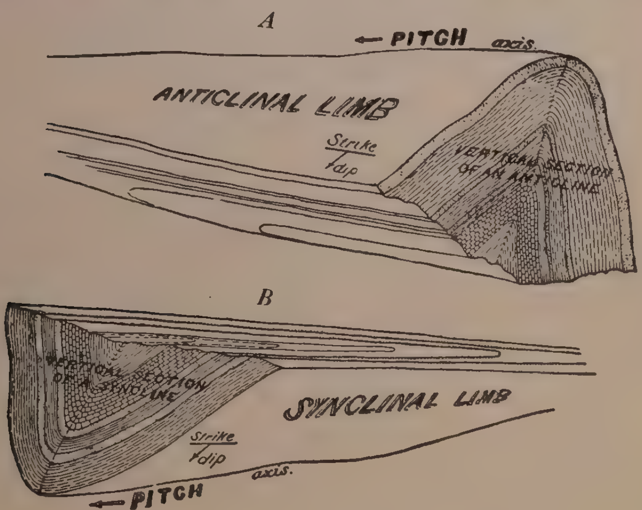

Fig. 61. Parts of folds. After Willis.

measured at right angles to the strike. If all rock layers were flat there would have been no occasion for these terms. Deviation from flatness is commonly, though not always, the result of folding, and so far as this is true *strike* and *dip* describe elements of folds. This may seem like emphasizing the obvious; but in the definitions of strike and dip presented by elementary textbooks there is often a concealed assumption of the relation of these observations to folding, which the elementary student does not discover until he starts field work. Within the writer's experience there have been many cases

where beginners have faithfully recorded strikes and dips over an area, apparently with the idea that these were the ultimate facts required, and failed to recognize the simple fact that a determination of strike and dip is merely a step in the elucidation of folding and other structures.

The *axial plane* of a fold is a plane which intersects the crest or trough in such a manner that the limbs or sides of the fold are more or less symmetrically arranged with reference to it. Notwithstanding its name, it is sometimes curved, and

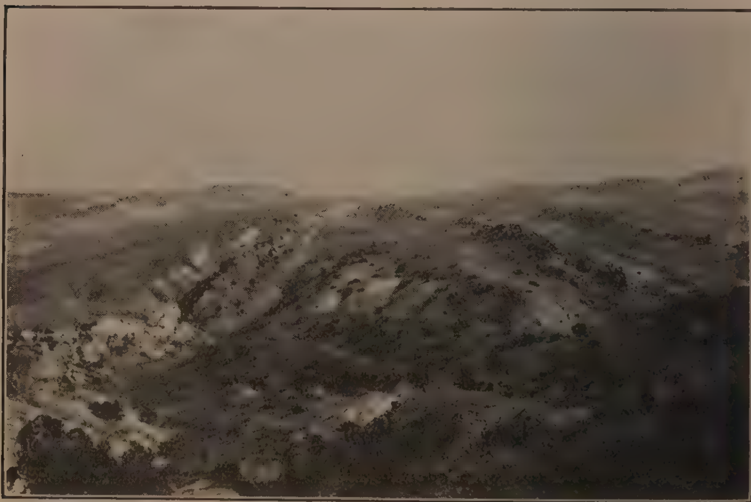

FIG. 62. Surface expression of an anticline looking along its axis in the direction of pitch. Between Bonanza and Tensleep, Wyo. After Lupton, U. S. Geol. Survey.

it may have any attitude from vertical to horizontal. The intersection of the axial plane with the crest or trough of a fold is the *axial line*, *axis*, *crest line*, or *trough line*. The *pitch* of the fold is the inclination of the axial line to the horizontal. It is merely a special case of dip taken along the axis of the fold.

Where the pitch is zero, in other words where the axial line of the fold is horizontal, the axial line is a great circle of the

earth and should theoretically have no end. An exactly horizontal axial line is a very rare occurrence. Where it is horizontal, the direction of the pitch and the strike of the beds in any part of the fold are parallel. Where the pitch is not zero the direction of the axial line and the strike are not parallel, except in closely compressed folds with vertical limbs.

The fact that a fold has pitch is ordinarily assumed to mean that its axial line is crossfolded, that is, that there has been longitudinal as well as transverse shortening. This is undoubtedly true in some cases, but in others pitch can exist without longitudinal shortening. The shortening suggested by the vertical curvature of the axial line may be more or less compensated for by tension cracks (see Fig. 77).

The attitude of the axial plane and the pitch of the fold are the most significant observations that can be made. They tell more about the three-dimensional position of the fold than do strike and dip observations. The main purpose of strike and dip observations is to get at the larger features of folds expressed in terms of axial plane and pitch. Strike and dip obviously vary on different parts of a fold; there must be plenty of observations more or less well distributed over the fold to get a true picture of it. In folds of simple outline a few well-distributed strike and dip observations may suffice. In much contorted folds strike and dip vary so widely and sharply that a thousand observations may not be sufficient for the purpose. It is often possible in these cases to confine observations solely to axial plane and pitch, — thus shortening the procedure necessary to an understanding of the major features of the fold. Efficiency is gained in structural mapping by going most directly for the ultimate structure.

Kinds of Folds

A *simple* fold is a single bend or flexure without minor crenulations. A *composite* fold is a major fold with minor crenulations superposed on it, with axes more or less parallel to the

axis of the major fold. *Major fold* and *minor fold* are terms expressing relative size. A major fold may have minor folds on its limbs; on these in turn still smaller folds may be superposed, and so on. The terms major and minor express the relation between any successive members of this sequence. The terms are also applied to independent folds not in such a sequence. Where minor folds are superposed on major folds, the axes of the minor folds have a tendency to be parallel to those of the major folds, but may also be in other directions.

The terms *simple* and *composite* ordinarily apply to two-dimensional cross-sections of folds, and there is an underlying assumption that the axial lines of the folds, extending in the third dimension, are approximately straight. Commonly, however, the axial lines themselves are not straight but are folded. The fold then becomes a *complex fold*. Most folds are really complex, and hence the terms *simple* and *composite* may be considered essentially as descriptions of complex folds as they appear in cross-section. It is not always easy in discussing folds to discriminate clearly between a consideration of two dimensions and of three dimensions, and hence the use of the terms "simple," "composite," and "complex" is in practice frequently loose. The terms are useful in keeping clearly before us the desirability of discrimination between two-dimensional and three-dimensional treatment of folds.

Anticline and *syncline* refer respectively to the arch and trough of a simple fold. As an anticline may be recumbent or overturned, it is more exact to define it as a fold in which the originally upper layers or parts are on the convex side. *Anticlinorium* and *synclinorium* refer to composite or complex arches and troughs.

Each of these kinds of folds may be further classed as *upright, inclined, overturned,* or *recumbent,* depending upon the attitude of its axial plane. No further definitions of these terms seem necessary. Where the limbs of a fold are parallel, it is called *isoclinal*. Where the axial planes of the minor folds of an anticlinorium converge downward, the fold is called by

Van Hise[1] a *normal anticlinorium;* a *fan fold* is a special case of this (Fig. 63). If they converge upward it is an *abnormal anticlinorium; roof structure* is a special case of this (Fig. 64).

A similar classification may be made for synclinoria; if the

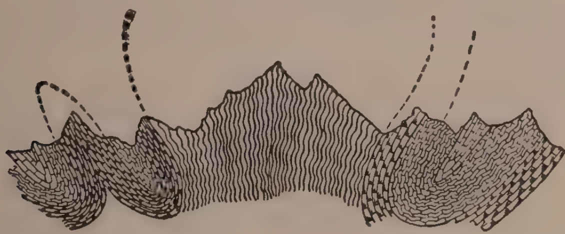

FIG. 63. Generalized fan fold or normal anticlinorium of central massif of the Alps. After Heim.

axial planes of the minor folds converge upward, the fold is normal; if they diverge, abnormal.

A *monocline* is a flexure in a straight bed, in the form of a double curve; the bed suddenly changes its dip and then resumes its dip. The term monocline or monoclinal dip is also applied to any sequence of beds which have the same

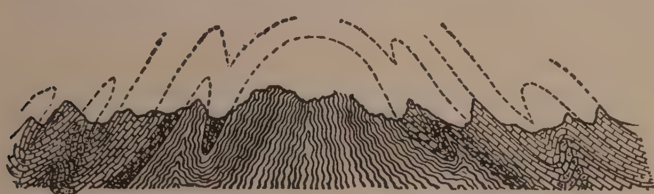

FIG. 64. Generalized section of roof structure or abnormal anticlinorium of the central massif of the Alps. After Heim.

attitude over a large area, even though this sequence be but part of a complete fold when a greater area is considered. Sometimes, also, it is convenient to describe a structure as monoclinal before it is known whether or not the structure is really isoclinal. A repetition of isoclinal folds, the crests of

[1] Van Hise, C. R., Principles of North American pre-Cambrian geology: 16th Ann. Rept., U. S. Geol. Survey, pt. 1, 1896, pp. 608–612.

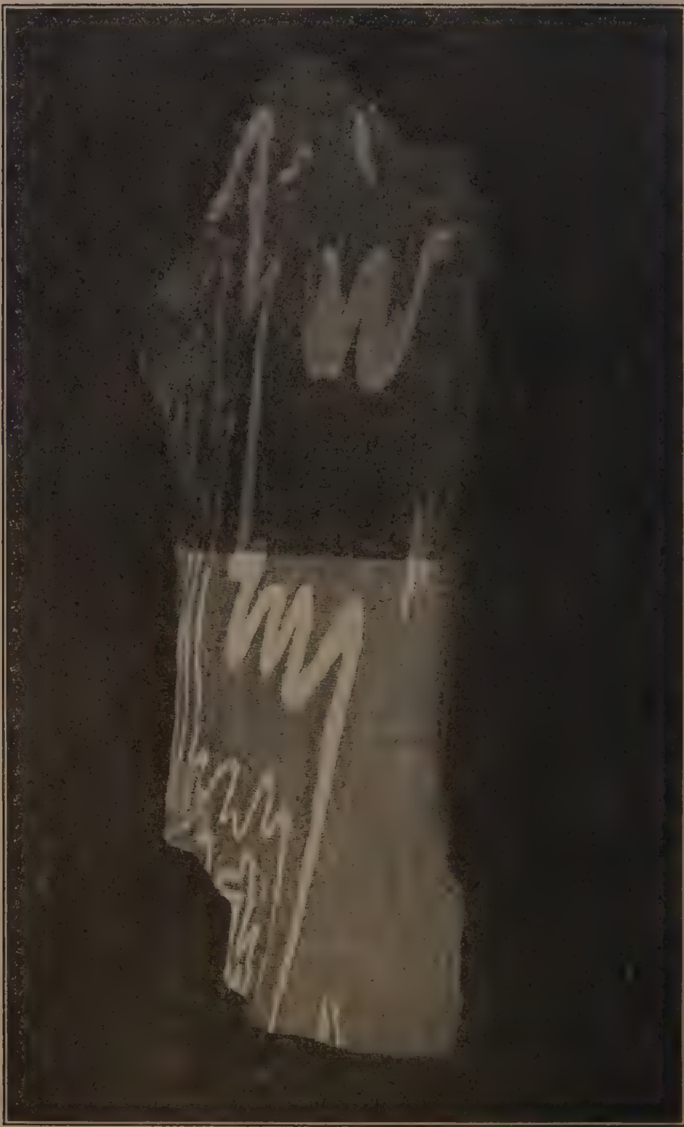

FIG. 65. Illustrating incompetent folds developed by rock flowage in gneiss. After Van Hise.

which have been eroded away, will yield a series of uniformly dipping beds in which the fact of monoclinal dip is apparent, but in which the detection of isoclines is slow and difficult.

Domes and *basins* are terms which explain themselves. They are said to have a *quaquaversal dip*, which means that the beds dip uniformly from or toward the center.

FIG. 66. Folded schist from Alaska. Folds are "similar" but the sharpness of the bends involves a minimum of distortion of the beds.

Drag folds are folds formed by the drag of one bed over another, as an incident of either major folding or faulting. Most folds are probably *drag* folds on a larger or smaller scale.

Folds in *competent* beds are on the whole likely to be of larger amplitude, more simple outline, and less sharply flexed than folds in less competent beds. Massive quartzite and dolomite formations often show, simple, well-curved outlines,

FIG. 67. Illustrating the artificial development of a fold. After Willis. The fold begins to develop at points of initial irregularity in the beds (initial dip) near the point of application of force. The heavy layers rise in simple, competent, parallel folds, the soft layers in composite, incompetent, similar folds. When the stronger layers have risen to the limit of their competency they buckle, developing composite outlines and to that extent taking on characteristics of incompetent folds. The force is applied at the right.

while softer shales and thin-bedded associated sediments are much more sharply and intricately folded. In *incompetent* beds a fold is likely to show sharper bending, the limbs being more or less straight. Rocks deformed by fracture on the

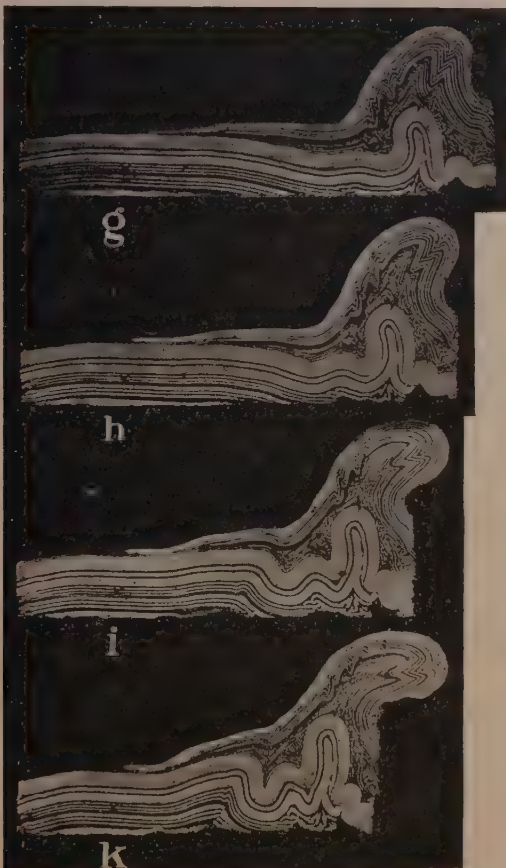

FIG. 67 (continued)

whole have been relatively competent, and rocks deformed by flowage have been incompetent. Hence the distinction between curved and angular folds corresponds somewhat to difference between folds accompanied by rock fracture and folds accom-

panied by rock flowage. It is to be remembered that competence is not only a function of the qualities of a rock bed itself, but also of the load under which it is deformed. Under extreme conditions of load a rock that is normally competent near the surface may become highly incompetent; and hence it is that in exceptional cases, where the pressure has been extraordinarily high, even strong rocks like quartzite and dolomite may exhibit sharply angular and intricate folding.

The use of the terms *competent* and *incompetent* respectively for the behavior of beds in folding requires some further explanation. Willis' experiments on the mechanics of Appalachian structure [1] showed that the thicker, more competent wax layers rise in simple outline under given conditions of pressure and load until they are unable to lift the load further. Then they crumple and, in crumpling, thicken, enabling them to lift the load higher. Thus composite folds are really indications of incompetence. Simple folds are more characteristic of conditions favoring fracture; the bed is able to lift itself without interior readjustment, and without crumpling; it is competent. All folds represent a yielding to pressure. In that sense all are incompetent, and it might be better to speak of them all in terms of degrees of incompetency. There is likely, however, to be little confusion in following Willis in the use of the two terms competent and incompetent.

Our field of observation is practically confined to a zone of combined fracture and flowage (see Chapter II), and hence to folds representing some combination of the two types of deformation. The folds described as typical of fracture and of flow may be regarded as the limiting cases between which all folds may be classified.

When folds are examined in detail it will be noted that some beds are thinned along the limbs and thickened along the crests and troughs. These are usually the softer, more incompetent beds, and they are likely to possess rock cleavage

[1] Willis, Bailey, Mechanics of Appalachian structure: 13th Ann. Rept. U. S. Geol. Survey, pt. 2, 1893, pp. 241–253.

or other evidence of flowage. Beds of a more competent type usually do not show the thickening and thinning; their thickness remains uniform. These beds are likely to show fracture. The movement is largely concentrated in slipping between the beds, whereas the thickening and thinning of the incompetent beds involve interior movement throughout. Few series of

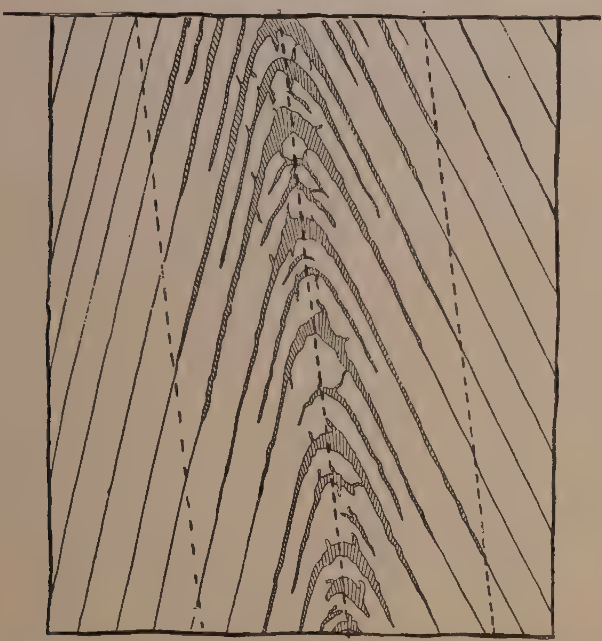

FIG. 68. Section of sharp fold in gold-bearing slates of Nova Scotia, showing saddle reefs of quartz. After Faribault.

beds are so homogeneous that they behave uniformly. Commonly there is a combination of the two types of folding, but one or the other may dominate, depending on whether the beds on the whole are more or less competent. The stronger layers in a series of heterogeneous beds tend to separate along axial lines when folded, one layer lifting itself off the other on the arching principle. A space between may remain as a void, but very commonly the softer layers flow from the limbs

toward this void and fill it, — or the space may be filled by vein material, constituting "saddle reefs," as in the gold-bearing slates of Nova Scotia and the Bendigo reefs of Australia (see Figs. 68, 69 and 70).

The forms of folds in these two types of deformation are distinctive and it is convenient to give names to them. Where

FIG. 69. Folding of brittle and soft layers contrasted in jasper. The broken, dark layers are chert, the light layers are iron oxide.

the beds are not thickened and thinned, the sides remaining parallel, the curvature in successive layers is more or less concentric. With increasing distance from the center the curvature is less. This type of folding is called *parallel* or *concentric folding*. The shortening of a bed away from the concentrum, through any unit of distance, is obviously less

than the shortening of the bed by curvature near the concentrum. This difference in the amount of shortening implies differential movement between the beds, which is an observed characteristic of this type. The lessening of the curvature away from the concentrum is often expressed in another way; a fold is said to die out. At an infinite distance from the

FIG. 70. Folding of brittle and soft layers contrasted in jasper. Note the tension cracks in the brittle layers.

center the curvature is so slight that the bed approaches straightness. By following along the axial plane of any fold in competent beds, change in curvature is to be seen, due to this tendency for the curvatures to be concentric. Folds of this type, as noted above, are the result of deformation principally by rock fracture.

Where the beds are thickened and thinned of course their sides do not remain parallel. They are not concentric. The curvature of the beds does not change, but is the same in one

bed as in another. There is no dying out of folds and no differential movement. This type of folding is called *similar*. Folds of this type are the result of deformation principally by rock flowage.

It is not easy to remember and contrast terms so alike in general meaning as *similar* and *parallel*. Students who try to memorize these terms, without getting first a clear mental picture of the fundamental differences, usually have difficulty in using them. For this reason there might be some advantages in using the term *concentric folds* in place of *parallel folds*.

In field work the discrimination of competent and incompetent folds, or of concentric and similar folds, is highly useful. If, for instance, rocks have in general a tendency toward the concentric type of folding, the investigator is led to look for the dying out of the fold, evidence of considerable differential movement, absence of thickening and thinning — all of which must be kept in mind in assembling data and drawing maps and sections. Without this conscious attempt to classify the fold it is a common experience that the geologist's representation of the fold fails to express its essential characteristics. Especially is this important in detailed analysis of structures affecting an ore deposit, where the assumption of uniformity in type of folding may lead to faulty inferences as to distribution of beds even a few feet beyond the field of direct observation. For instance, in explorations for iron ore in the great slate areas of Lake Superior, it soon becomes apparent that the slate has been folded under conditions of flowage. The observer is therefore justified in concluding that the folding is probably close and complex, that there was much thickening and thinning of the beds, that the folds were largely of a similar type, not dying out above or below. The application of these principles has been of great aid in interpreting fragmentary records brought up from the drill holes or in outcrops. In the Marquette district of Michigan, where there are beds of quartzite interbedded with softer slate and iron

formation, it has been possible by the application of these principles to correlate some of the simpler and broader structures of the quartzites with the closer, much more complex, and quite different folds of the softer beds. In making any satisfactory estimate of the thickness of folded beds the first

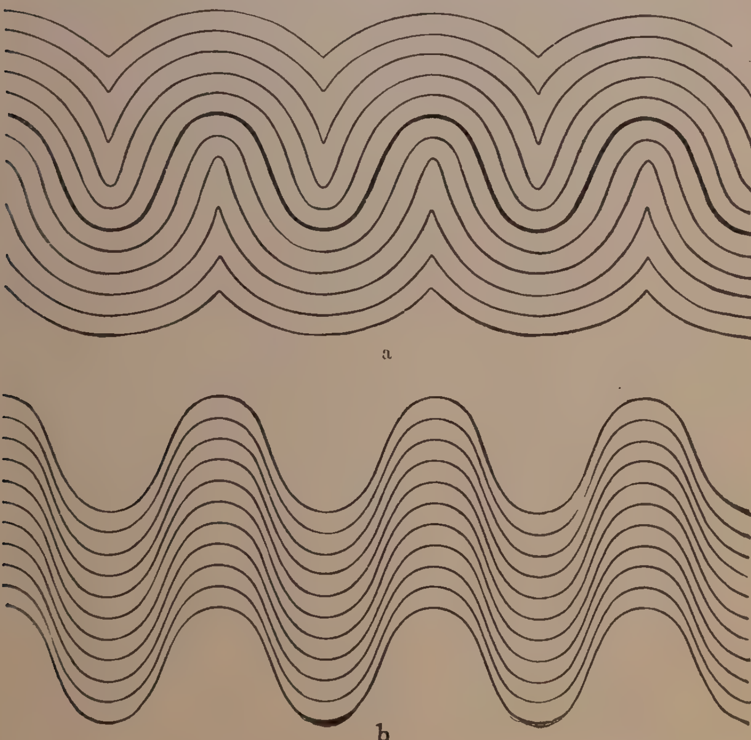

FIG. 71. Figures illustrating (*a*) ideal parallel or concentric and (*b*) ideal similar folds. After Van Hise.

question to be settled is the degree to which the folds are characteristically those accompanying rock flowage, and therefore to what extent the beds are likely to be thickened or thinned.

Folds have such a great variety of forms and origin that

names applied to them cannot for the most part be regarded as exclusive or final. By names we merely emphasize this or that characteristic which for the moment seems to us essential. A fold may be, for instance, an anticline or anticlinorium and at the same time be overturned, recumbent, or isoclinal, concentric or similar, or a drag fold. A fold may be simple in one cross-section and composite in another, or may be upright in one cross-section and overturned in another. Still more complexity is introduced by the fact that rock beds on the whole are not homogeneous, and that different layers or parts do not act the same when deformed. Certain layers may fold in one way and others above and below or between fold in another way. The student should not expect to find many picture-book illustrations of fold types free from complication with other types. Let him crumple up a folio of paper of different weights, ranging from cardboard to tissue, and attempt to name the fold as a whole, and then to name it as it would appear in different areas and cross-sections. Complexity of folding may be so great, as it often is in nature, that rigidly accurate classification and naming are practically an impossibility.

Control of Flow Cleavage, Fracture Cleavage, Jointing, and Drag Folds by the Differential Slipping Between Beds on Limbs of Folds

One of the most significant features of folding from the standpoint of structural diagnosis is the fact that it is impossible to fold a series of conformable layers without causing a slipping between them, — as may be seen when a book of paper leaves is flexed. In an anticline the upper beds move over lower beds toward the anticlinal axis; in a syncline the lower beds move past higher beds toward the synclinal axis. Not only is there movement between the beds, but there is similar movement within the beds themselves, more or less in proportion to their weakness. In a set of heterogeneous beds evidences of movement are likely to be conspicuous only in a few weak beds.

The structural results of such movement are expressed as flow cleavage, fracture cleavage, jointing, and drag folds, which have definite relations, noted below, to the elements of folding and serve as diagnostic criteria in the study of folds.

Drag folds in relation to major folding. Minor folds may result from differential movement between beds on the limbs of major folds. Such folds are ordinarily known as *drag folds.* Drag folds may also result from slipping along faults

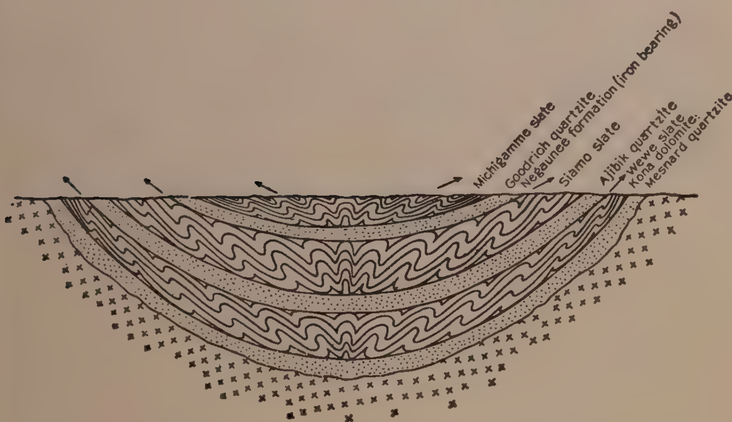

FIG. 72. Figure showing differential movement between competent beds on limbs of a fold with the development of minor drag folds between them. Marquette district, Mich.

(see Figs. 31 and 32). In the folding of heterogeneous rock layers the stronger, more competent layers are likely to show broad, simply flexed outlines, and the softer, less competent layers to show more intricate minor foldings of a drag type. Drag folds do not always develop under these conditions; the deformation may be by jointing or flow cleavage. The axial planes of the drag folds tend to be more or less parallel to the axial plane of the major fold whether they are on the limbs or axis of the major fold, but may also vary a few degrees from this attitude, in which case the variation is the basis for special names for the structure. Where, on an anticline, the

axial planes of the minor folds on the limbs diverge upward, the structure is called a *normal anticlinorium.* Where they converge upward toward the axial plane of the major fold it is called an *abnormal anticlinorium.* The difference may be merely one of degree of rotation of the axial planes of the drag folds by the differential movement between controlling beds on the limbs. Where there is a large amount of differential movement concentrated in any given bed the rotation is greater and the fold is more likely to become abnormal. Where the movement is not sufficiently great or concentrated the rotation may not be enough to produce the abnormal attitude, and the normal fold results.

The writer is inclined to emphasize the essential parallelism of the axial planes of minor and major folds as the most common and important relationship. When the observer finds a small fold in a weak bed, suspected to be of drag type, he will find it reasonably safe to infer that its axial plane is substantially parallel to the axial plane of the major fold, and this inference may be very helpful in following up the larger and yet unseen structure. The determination later that the minor folds have either normal or abnormal relation to the major fold is a refinement which will not seriously disturb his earlier assumption. In such cases the variation of the axial planes from parallelism to the axial planes of the major folding is usually slight.

An essential characteristic of drag folds is their confinement to certain layers, while adjacent layers do not have them at all or have them less well developed. It is just as if a sheet of tissue paper on a desk were crumpled by sliding a book across it. The deformed layer is actually shortened, whereas the adjacent layers are neither lengthened nor shortened; they are merely slipped one past the other. It follows that the folding and thickening of the weak bed in one place must be complemented by its thinning or parting in another. Thinning of beds is a process which usually occurs on the limbs of major folds, and thickening along their crests and troughs;

therefore drag folding is likely to be more conspicuous near the axes of the major folds than on the limbs. This fact is to be kept in mind in estimating the shortening of an area by folding. Another essential characteristic of drag folds is that, so far as they have any persistence, it is mainly in the direction of their axial lines. Not only are they confined to certain weak beds, but they are likely to follow one linear direction within the bed, and other parts of the bed may be undeformed. This tendency to linear persistency is not strong, for the folds often die out along their axial lines, or may be replaced by parallel offset folds.

As folds usually have a pitch, the axial lines of minor drag folds, when projected to a horizontal plane, usually vary a few degrees from the strike of beds on the limbs of major folds unless the major folding is so close as to make the limbs parallel. This is illustrated by folds in the iron formation of the Menominee district of Michigan (Fig. 73). At one place the iron formation dips 70° N. and strikes N. 70° W. The pitch of the minor folds is 30° in a direction N. 65° W. As the pitch carries these folds down they are carried northward down the dip of the beds. Hence in this case there is a divergence of 5° between the surface projection of the axial line of the minor fold and the strike of the bedding, which is a fact of some commercial significance in the exploration for ore, in view of the fact that the ore follows the pitch rather than the strike.

Drag folds afford a means of determining which is the top and which is the bottom of a bed on the limb of a major fold. If in an isolated outcrop of vertical beds it is apparent from the minor drag folds that the left-hand side has moved up with reference to the right-hand side, the inference is that the ledge is a part of the left limb of an anticline. If so, the top of the beds is to the left. The same inferences are to be drawn from fracture cleavage, flow cleavage, and certain joints caused by differential movements in folding.

In general, the relations of drag folds to major folds, above

described, furnish highly useful evidence in structural field work. Adherence to the simple plan of watching for indications of differential movement as expressed by drag folding often leads to surprising results. In districts where folding is extremely intricate, the observer may be at first bewildered by the apparently chaotic structural condition; but it may be found that the minor contortions indicate some general

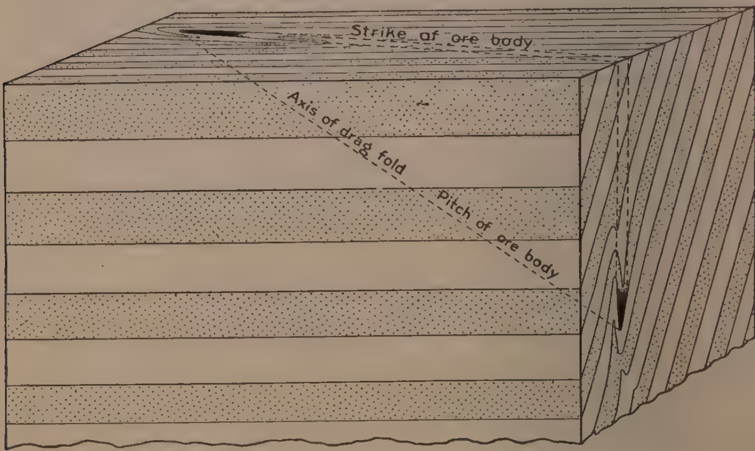

FIG. 73. To illustrate divergence in strike and pitch. After Mead.

drag between competent layers, and if so, this gives a notion as to the position of the next larger unit of structure. This leads to a study of still larger units, and so on. In the Marquette district of Michigan the slate beds are folded in the manner to be expected from the control of the harder quartzite layers of the Marquette synclinorium. Understanding this relation, the intricate outlines of the slate folds come to express some order, which may be satisfactorily correlated with the simple outlines of the quartzite folds. In turn the Marquette synclinorium as a whole may be regarded as a minor fold showing differential movement upon the limb of the major Lake Superior synclinorium.

The principle of the control of minor by major folds affords

the most reasonable hope of working out successfully the complex structures of the old Archean or Basement Complex, which heretofore have been regarded as almost inexplicable. Although on casual inspection the folds in any ledge show an apparently great complexity, when examined with reference to the differential movement the general structure becomes more perceptible, and it is possible to infer some of the relations of the major folding.

The "decke" or "nappe" structure of the Alps, illustrated by Fig. 31, is a series of great overthrust folds, associated with faults, with nearly parallel and horizontal axial planes, which are probably to be regarded on a large scale as "drag folds" resulting from the horizontal shearing of some formerly existing competent rocks over the Alpine area. The great Alpine fan folds of the type so well known through the writings and sections of Heim [1] and others are now being interpreted as nappes by Schardt, Lugeon [2] and others. The actually observed structures seem to permit of connections on cross-sections drawn to correspond to either hypothesis, and it is probably uncertain in some cases which interpretation is the correct one.

Flow cleavage in relation to folding. Flow cleavage may occur in the weaker folded beds while the stronger beds are fractured. Not all weak beds show flow cleavage; some may have flowed without producing flow cleavage, or they may have been under such conditions of load that they fractured rather than flowed. Mention has already been made of the fact that some folds are so gentle that no one small part of them exhibits either flow or fracture on a scale ordinarily identifiable. Cleavage also may be present in rocks which have not been folded, as in the region of certain thrust faults, and perhaps also in cases of load metamorphism (see p. 129).

[1] Heim, Alb., Untersuchungen über den Mechanismus der Gebirgsbildung, Basel, 1878.

[2] See: Der Bau der Schweizeralpen, by Alb. Heim: Neujahrsblatt der Naturforschenden Gesellschaft in Zürich auf das Jahr 1908, 110 Stück.

However, in by far the most common case, cleavage is the result of distortion of beds or other structural units during folding, and its very presence is strong presumptive evidence of folding. The cleavage structure indicates that the rock has yielded to deforming forces throughout, and this yielding usually involves contortions in the layers. Even where these contortions are not present, the cleavable bed as a whole may be folded.

The flow cleavage lies approximately parallel to the axial planes of both the major and minor folds. It follows that it is inclined to bedding, and that its inclination varies in direction and degree on different parts of the folds. Figure 47 indicates the typical relationship. At the crests and troughs only is the cleavage likely to be normal to the bedding. Here there is no differential movement.

As most folds have a pitch it follows that the horizontal plan will look very much like the vertical section, with cleavage inclined to bedding on the limbs and normal to bedding on the crests or troughs. Differential movement between the beds may be determined from the angle of the cleavage with bedding, just as it is in cross-section. For instance, an east-west vertical cleavage may be crossed by vertical bedding running northeast. The inference is that the beds to the southeast have moved northeast with reference to the beds to the northwest, and that toward the northeast the beds will be found to turn northwesterly about an axial line of the fold, while to the southwest they will turn in to the southeast.

With cleavage alone observed, we may infer that there is probably folding, and that the strike and dip of the bedding are other than that of the cleavage. Where a strong regional cleavage is emphasized at the surface by weathering, it is easy to be misled into thinking that this marks the general strike of beds, especially in a closely compressed district where the angle between beds and cleavage is not a large one. It is well to remember that the cleavage direction is usually not the direction of strike, and that in tracing beds

through the complex the probability is that they will connect in directions other than along the cleavage.

If, in addition, the bedding can be observed, its angular relation to the cleavage shows in what part of the fold the observation is taken, — as brought out in the example above.

If the bedding is vertical, the inclination of the cleavage to it will show which side has moved upward, and therefore which is the top.

Cases of overturned folds can be detected by noting the relation of cleavage to the bedding; thus, if the topmost of a series of inclined beds is shown by cleavage to have moved downward with reference to an underlying bed, the relationship is abnormal, and warrants looking for evidence of overturn. In the normal relationship the cleavage is likely to be steeper than the bedding, but when the bed is overturned the reverse is true. Evidence of this kind, of course, is checked by observation of drag folds, of fracture cleavage, and by the evidences of top and bottom beds furnished by the structures of primary deposition, — such as ripple marks, false bedding, and gradations in texture (see pp. 187–189).

Still further, cleavage being parallel to the axial plane of the fold, it follows that the trace of the bedding on any cleavage surface gives approximately the pitch of the fold. This is exactly true for the particular cleavage plane which coincides with the axial plane, but it is approximately true also for any other parallel cleavage plane. If the student will fold a soft clay bed and section it in several places parallel to the axial plane, he will see that the intersections of the bedding with the planes of the several sections will give the direction of pitch, and usually the approximate degree or angle of pitch, although the angles thus observed may vary with the position of the section.

In composite folds of heterogeneous strata the axial planes of the minor folds may not be entirely parallel, and therefore the cleavage is not entirely parallel throughout the major

fold. In general, however, even in these cases there is an approximation to parallelism.

It is a noteworthy fact that the strike and dip of cleavage in any deformed area are much more uniform than the strike and dip of bedding. The cleavage may be parallel over a large area whereas the beds are much folded. This merely expresses in another way the general parallelism of cleavage to axial planes; for on the whole, in any area sufficiently compressed to make a cleavage, the folds are likely to have a considerable degree of parallelism of their axial planes.

With this normal relationship of cleavage to folds in mind it becomes possible to use cleavage in a large variety of ways as an aid in integrating the structure as a whole. No matter how fragmentary the evidence, some useful inference may be drawn. There are many combinations of the several factors in the problem, and to develop in detail all of the inferences from the various possible combinations would unduly extend this text. It is not a matter of hypothesis; it is an empirical relationship established by wide observation. Only field experience can give facility in using this method effectively.

For illustrative purposes we present one extreme case of inductive reasoning from cleavage in drill cores brought up from depth. It involves some interesting problems of descriptive geometry.[1] If the cleavage is found to be vertical, it may be assumed that the axial planes of the folds are vertical; if inclined the axial planes of the folds are inclined. Drilling may be done to determine the exact strike and dip of cleavage, in which case the axial planes of folds are ascertained accordingly. If cleavage is found to be normal to the bedding, the trough or axis of a fold is assumed; if inclined to the bedding, a limb of a fold. If cleavage is steeper than bedding,

[1] Mead has described the methods applied to the interpretation of bedding shown in drill cores, and they will not be repeated here. When cleavage is shown in addition to bedding, a long series of additional inferences may be drawn which are based on the relations of cleavage to folding above described. Mead, W. J., Determination of attitude of concealed bedded formations by diamond drilling: Econ. Geol., vol. 16, 1921, pp. 37–47.

normal, upright folds are assumed; if flatter than bedding, there is presumption of overturn. With the direction of cleavage known, the trace of bedding on the cleavage surfaces, even on the small pieces of drill core, will determine pitch.

Jointing and fracture cleavage as evidences of differential movement between beds in folding. Differential movement between beds develops one set of shearing planes parallel to the bedding and another at an angle, more nearly parallel to the axial plane of the fold than to the bedding (see Fig. 6). Joints or fracture cleavage form along these planes. They may be curved or S-shaped. Also they are likely to be confined to certain beds, and to offset along the bedding planes in passing to different strata on either side. Given, then, joints or fracture cleavage thus obviously related to folding, it is possible to determine the differential movement and to ascertain what part of the fold is under observation. In the Baraboo district of Wisconsin, northward-dipping beds of quartzite are cut by two sets of closely spaced joints or fracture cleavage, one set parallel to the bedding, and another set crossing the bedding and dipping northward (see Figs. 57 and 58). It is clear in this instance that the upper beds have moved southward with reference to the lower beds. This corresponds to the requirements of a position on the south limb of a syncline.

With the differential movement indicated in this way, the top and bottom of beds may be inferred, as in the case of drag folds and flow cleavage.

Determination of Top and Bottom of Sedimentary Beds in a Folded Area

The relations of cleavage, joints, and minor drag folds to major folds, discussed above, help to determine which is top and which is bottom of beds. In addition there are primary structures of beds which may be used to advantage, particularly (*a*) ripple marks, (*b*) false bedding, and (*c*) variations in coarseness of grain.

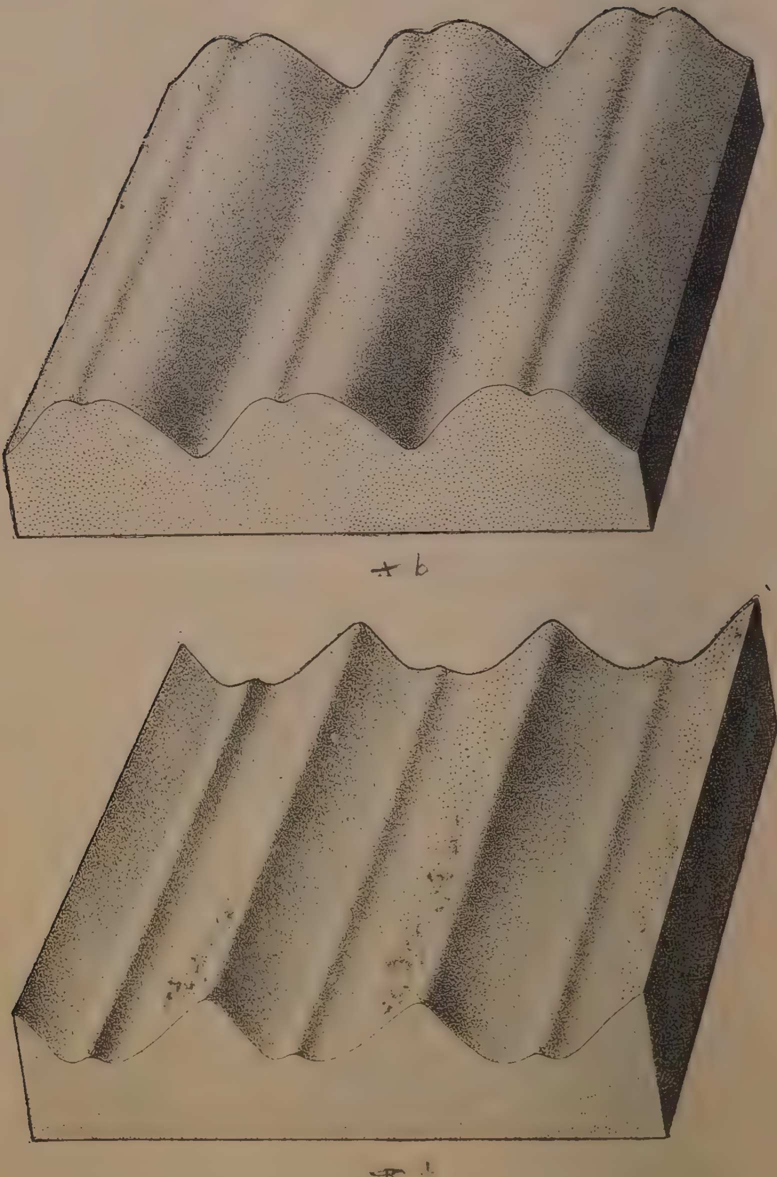

FIG. 74. Photograph of (*a*) ripple marks and (*b*) casts of ripple marks. After Van Hise.

(*a*) In Fig. 74 the normal ripple marks and their casts are indicated. It will be noted that in the normal ripple marks the crests are much sharper than the troughs, and that the troughs may have minor crests in them. When the beds are on edge or overturned, these facts enable one to tell which is top and which is bottom. Care must be used in applying this criterion, however, for it may only be used in the case of symmetrical ripple marks such as those illustrated, which are formed by wave oscillations in shallow waters. Ripple marks formed by currents of air and running water are of an asymmetrical type, in which the true ripples and the casts are so similar that they are usually of no value for determining top and bottom.

(*b*) In Fig. 75 it will be noted that the false bedding is abruptly cut off by overlying beds while it comes in contact with the lower beds by a tangential curve. If the outcrop shown in the photograph were turned on edge or overturned, there would still be no difficulty in determining which were the original top and bottom of the beds.

(*c*) It is common to find a diminution in coarseness of beds from the bottom toward the top. Even in microscopic sections this is apparent. A bed may start in abruptly with coarse sediments, these gradually becoming finer-grained above, and the next bed starting in again abruptly with coarser sediments. There is little difficulty in these cases, no matter what the folding, in determining the original top and bottom of the beds. This has been found especially useful in interpreting drill samples from folded rocks.

Various other criteria for determining top and bottom of beds are useful in certain cases. Unconformities, with or without basal conglomerates (Chapter XII), slight local disconformities due to contemporaneous erosion or scour of sediments during the process of deposition (p. 208), mud-cracks, fossil trails and tracks, and the position of certain types of fossil shells, may all be useful. The keen observer will often

FIG. 75. False bedding or cross bedding in sandstone. Dalles of the Wisconsin. After Salisbury and Atwood. Note that the false bedding joins the horizontal or main bedding below tangentially, while above it is sharply beveled by the overlying bed.

discover other criteria which in particular cases will give the solution of this important problem.

Forces Causing Folds

The forces causing folds are commonly orogenic disturbances which have exerted pressure from without; but metamorphic changes of all kinds, especially solution, involve volume changes which, acting in conjunction with gravity, may be translated into linear movements and result in folding (see pp. 224–225). Especially to be noted are the volume changes and slump in the drying and recrystallization of sediments.

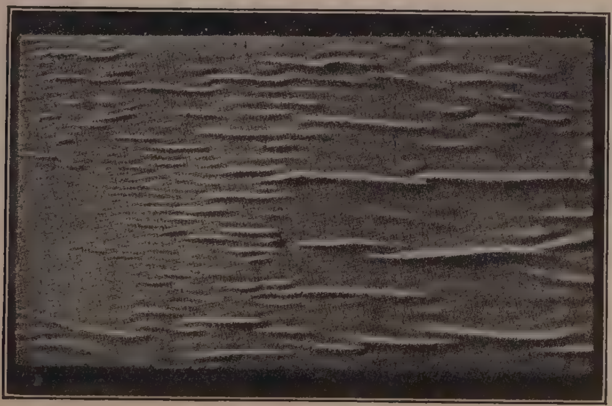

Fig. 76. Plaster of Paris model of folds produced by compression normal to the axial planes of the folds. After Mead.

Folding usually involves shortening in a direction normal to the axial plane, though to this there are exceptions noted below. It does not follow that the shortening is always caused by non-rotational stress applied normal to this plane. It may be accomplished in this way, but also by stresses inclined to the axial plane, i.e., by shearing or rotational stress, or by some combination of rotational and non-rotational stress (see pp. 18–19). Whether the stresses producing folds are applied normal to the axial planes or inclined to them, they

seem in most cases to be acting tangentially to the earth's surface and to cause tangential shortening. The fact should not be overlooked, however, that purely vertical movements, due to settling or to intrusion from below, may also cause folds by cross-bending, in which case there may be no shortening normal to their axial planes; there may even be extension. Where a normal fault passes along the strike into a monoclinal flexure, or into a flexure with more complex outlines, it is clear that there has probably been extension rather than compression.

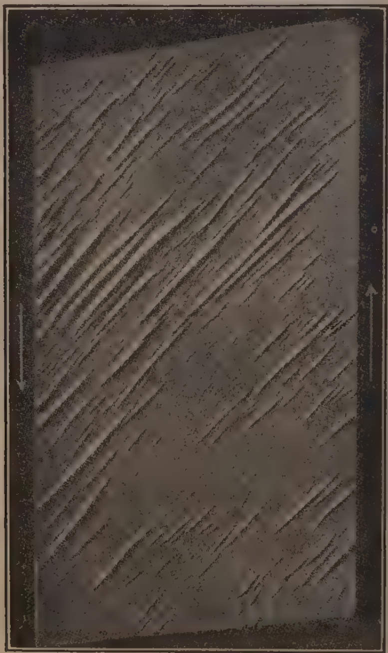

FIG. 77. Vertical view of reproduction in plaster of Paris of folds produced by shearing deformation. The direction of movement is indicated by arrows and the amount of deformation is shown by the shape of the block. Tension cracks cross the axial lines at right angles. After Mead.

While the forces causing folds as a whole are compressional,

within the fold itself they may become partly or wholly tensional after the bending has started. When a beam is flexed there is compression on the concave side and tension on the convex side (see Fig. 15). The surface separating zones of tension and compression is called the *neutral surface* or *neutral plane,* along which there is neither tension nor compression. The intersection of this surface with either side of the beam is called the *neutral line.* Within the elastic limit the position of the neutral surface may be definitely calculated; the neutral axis of any cross-section of the beam passes through the center of gravity of that section, provided that all of the forces applied to the beam are transverse. Engineers treat of the neutral plane only within this limit. Beyond the elastic limit, when failure begins, the position of the neutral surface involves more factors, and especially the relative strength of the material itself under tension as compared with compression. This is the situation we have to deal with in structural geology, where we are considering mainly failure beyond the elastic limit. The tensional strength of rocks is much less than the compressional strength, with the result that when an unloaded rock beam is flexed, the neutral plane, as soon as failure begins, migrates rapidly toward the concave side. Practically the entire beam is under tension and yields by fracture. On the other hand, where the mass is heavily loaded on all sides, as in a zone of rock flowage, all parts of it are under compression and the neutral plane may be considered as existing outside of the convex surface.[1]

In an intermediate stage the combination of load and resistance of material may cause a neutral plane within the mass itself, in which case the deformation is partly by tension on the convex side and partly by compression on the concave side. This latter case has been often presented in discussions

[1] Where the rock mass is flexed by pressures tangential to the earth's surface, also, the stresses are more comparable to those in a column than in a beam, — involving axial compression throughout in addition to the stresses of cross-bending. For theoretical discussion of this subject, see any standard textbook in mechanics of materials.

of structural geology in support of the conclusion that tension joints commonly develop on the convex side of folds. While

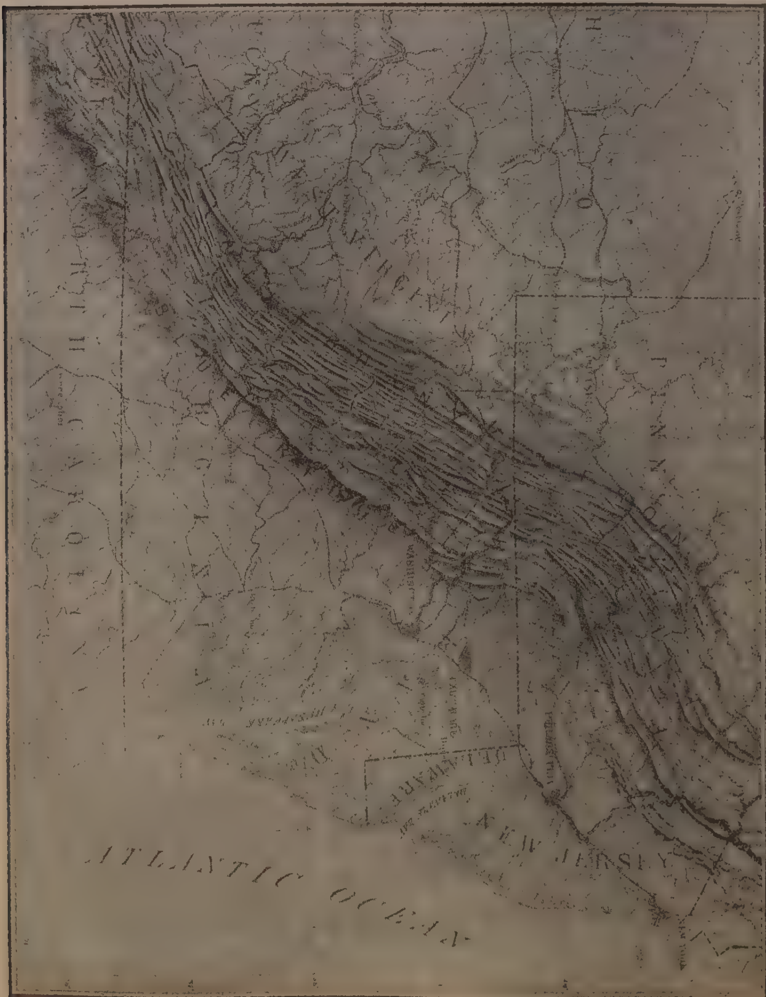

Fig. 78. Photograph of relief map of the Southern Appalachians, showing distribution of folds. After Willis. Compare with Fig. 77.

this condition undoubtedly exists in some folds, field observation seems to indicate that after all it is exceptional. Belief in its importance seems to be based on an assumed position of the neutral plane which does not hold good for two common limiting cases, namely, deformation near the surface mainly by fracture and deformation deep below the surface mainly by flowage. When we take into account the fact that deformed masses are likely to be heterogeneous in their strength and degree of loading, it is apparent that the simple assumption of the existence and position of a single neutral plane needs much qualification.

Folded regions often give evidence of repeated movements; the folds now observed are the result of more than one deformation. The repetition of movements tends to increase the complexity of structure, in which case there may be departure from some of the more simple structural relations, characteristic of a single movement, to which the discussion in this chapter is mainly confined. In most cases, however, later movements seem to follow zones of earlier movement, intensifying the structures first formed and not materially obscuring them. Even in much-folded terranes like the pre-Cambrian, which have been subjected to several periods of folding, the relationships of the structures and the folds still remain much the same as they would appear as the result of a single homogeneous movement.

Localization of Folds

(*a*) It has been shown experimentally by Willis [1] that folds tend to form near the point of application of the deforming force, unless the rocks are sufficiently rigid to transmit the thrust forward to some weaker zone. (*b*) Willis has also shown that slight irregularities in the bedding, such as might be formed during sedimentation, and which he calls

[1] Willis, Bailey, Mechanics of Appalachian structure: 13th Ann. Rept., U. S. Geol. Survey, pt. 2, 1893, p. 247.

initial dip, tend to localize a fold, even at some distance from the point of application of the force. (*c*) The uplift of a fold thickens and strengthens the mass, increases the local load, and tends to depress the beds immediately beyond it, creating an irregularity or "initial dip" which localizes another fold. The first fold rises to such a point that it becomes easier to develop a new fold in a more vulnerable place than to lift the old fold higher. (*d*) Dragging along faults sometimes causes folding. (*e*) Contact of rocks of unequal strength, for instance of granite and sediments, has been observed to localize folds, the massive granite serving as a buttress against which the weaker series is deformed. (*f*) Inherent weakness of rocks may localize folds. A shale is likely to be more folded than an adjacent quartzite. Initial dip or other irregularity may determine at what points in the shale the folds shall be localized, but the weakness of the shale as a whole as compared with the adjacent beds will favor the development of folds in the shale rather than in the quartzite. (*g*) Plutonic intrusion is responsible for much folding, and hence for localization of folds in place and time.

Folds formed in unconsolidated sediments during deposition are likely to be confined to thin horizons, though they may be extensive in these horizons.

Folds caused by leaching and slump usually owe their localization to conditions controlling the circulation of water. Thus in the Lake Superior region the leaching of quartz from iron formation follows well-marked water channels, determined by a variety of structural conditions.

Folds in gypsum beds caused by increase of volume during recrystallization are likewise confined to thin horizons but may be extensive within these horizons.

Coral reefs are commonly associated with an up-bending of strata around them and a down-bending below them.

Folds caused by the original deposition and slumping of sediments on a basement of irregular relief inherit their distribution from the topography of the old surface.

Depth of Folding

Within our zone of observation individual folds on the whole are not very persistent structures. They die out along the trend of their axial lines, and also upward and downward from the axes of closest folding. Folds of a similar type have more persistence than those of a parallel type. Zones of folding, on the other hand, may be persistent horizontally for thousands of miles, as in mountain chains, and the supposition is that such zones may be deeper than individual folds. In some places deep-lying rocks may be seen to be less folded than surficial rocks. Daly [1] finds in south-central British Columbia that the pre-Cambrian formations are much less folded than the overlying Carboniferous and Triassic rocks, indicating that a small depth of the earth shell has suffered strong folding in the post-Cambrian time. In other regions, such for instance as the Lake Superior pre-Cambrian, the older rocks are on the whole folded more than the upper ones. In general older formations are likely to be more folded than younger ones, because they have existed through more periods of folding.

It is sometimes assumed that the principal locus of folding is some distance below the surface, and that the surficial deformation is principally by fracturing without folding. This assumption does not accord with the evidence disclosed by the erosion of ancient structures, — where folds are found to be associated with both fracture and flowage, and are not more developed in lower than in higher horizons, except in so far as the lower are older and have been folded more times.

We ordinarily see only the surface part of a fold. In projecting any fold or series of folds into an unseen area below, we take into account the competence of the beds, whether the folding is of similar or concentric type, the relation to any great shear zone of known attitude, or to any intrusive; and in

[1] Daly, R. A., Pre-Cambrian formations in south-central British Columbia: Abstract, Bull. Geol. Soc. Am., vol. 23, 1912, p. 721.

this way we may sometimes arrive at a notion as to the persistence or lack of persistence of the fold. But there is another method.

By a simple geometric calculation it may be determined how much material has probably been affected to produce a given uplift by folding (see Fig. 79). In an illustrative case 100 miles of surface has been crowded into 75 miles. There has been an uplift of approximately a mile. Obviously the product of the linear uplift and the length of the shortened area, 1 mile x 75 miles, should equal the product of the

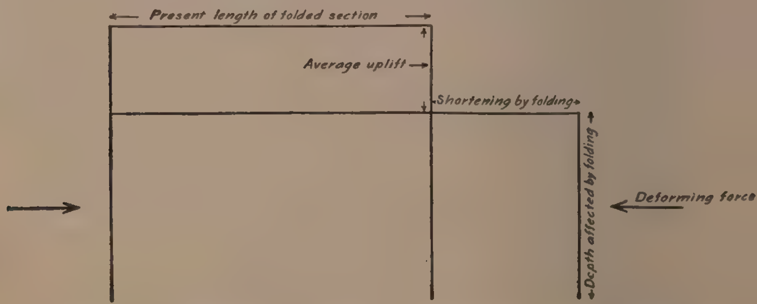

FIG. 79. Illustrating a method of determining depth affected by folds.

shortening, 25 miles, and the depth affected. By solving the equation, this depth is found to be 3 miles. This means that with a given elevation, the less close the folding (or faulting) and therefore the smaller the shortening, the greater the vertical depth involved in the deformation. In other words, moderate folding distributed through a thicker zone accomplishes the same amount of vertical bulging as closer folding of a thinner zone.

This method was suggested by T. C. Chamberlin,[1] and first applied by R. T. Chamberlin[2] to the Appalachian folds of central Pennsylvania. In a section from Harrisburg to Ty-

[1] Chamberlin, T. C., and Salisbury, R. D., Geology, vol. 2, 1906, pp. 125–126.

[2] Chamberlin, R. T., Appalachian folds of central Pennsylvania: Jour. Geol., vol. 18, 1910, pp. 228–251.

rone in Pennsylvania he found that shallower depths are affected on the two ends of the section, and greater depths toward the center (See Fig. 80). The shallowest deformation found is 5.7 miles. Making calculations for five sub-sections, he found a gradual increase in the depth affected toward the center of his section, which suggests that the deformed zone is bounded by planes dipping approximately 45° from the surface at either end of the section, and intersecting about 32 miles below the surface near the center; in other words, the deformed zone is wedge-shaped, with the flat side up. The intersection of these hypothetical planes at 45° with each other and with the earth's surface suggests to Chamberlin that they are really shearing planes developed by tangential shortening in the manner of fracture planes formed in a block under pressure, though within these limits the deformation may be by any combination of folding, flowage, and fracture.

A later application of the same method to a section across the Rocky Mountains, by R. T. Chamberlin,[1] indicates that the depth affected is at a maximum 107 miles. There is much less horizontal compression than in the Appalachians, and more vertical lifting and plateau-forming movement. Again the deepest deformation is near the center, suggesting that the folding involved a wedge-shaped mass.

Willis [2] applied the same method to the warping of a peneplain in the Cascade Mountains, and found the depth affected to range from 37.5 to 1500 miles. The entire mass is thought to have been shortened by flowage down to these depths, resulting in vertical uplifting.

It is to be noted that such calculations of the depth affected by folds are based on the assumption that the uplift bears a direct relationship to the crustal shortening, and the resulting figures cover only the uplift due to this cause. It is known

[1] Chamberlin, R. T., The building of the Colorado Rockies: Jour. Geol., vol. 27, 1919, pp. 145–164; 225–251.

[2] Willis, Bailey, Physiography and deformation of the Wenatchee-Chelan district, Cascade Range: Prof. Paper No. 19, U. S. Geol Survey, 1903, pp. 92–97.

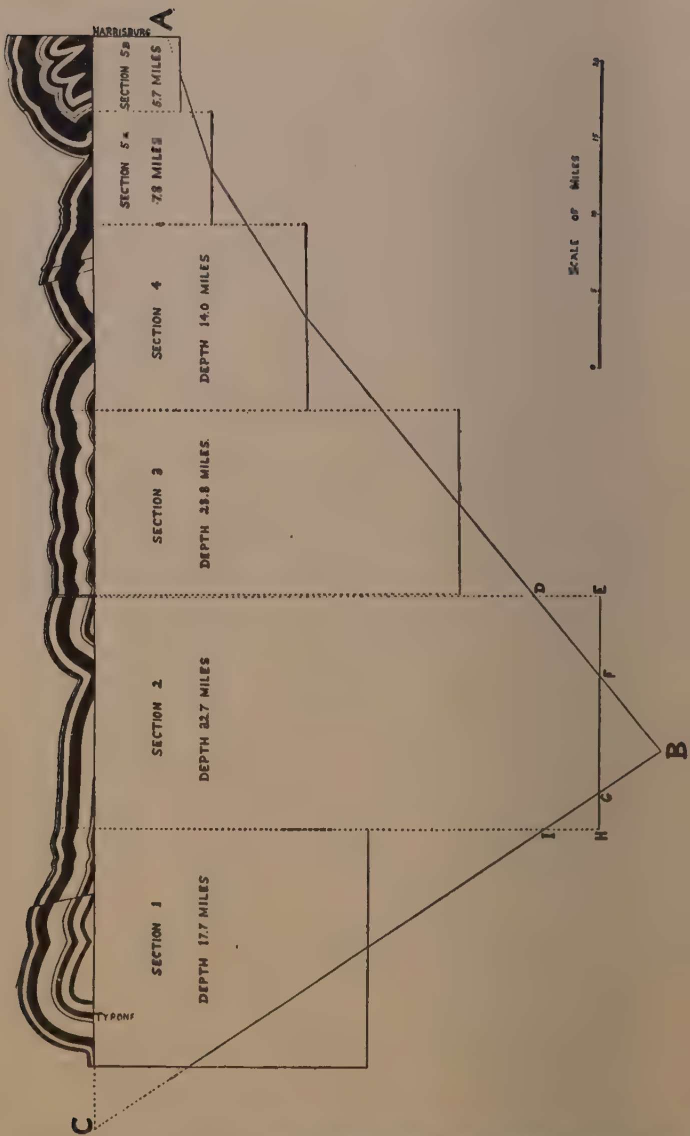

FIG. 80. Plot of the Tyrone-Harrisburg folded section representing the thickness of deformed shell beneath each of the six blocks as determined by the above methods of measurement. After R. T. Chamberlin. The lines AB and BC are drawn through the middle points of the bottom lines of each of these blocks, except Section 2, the apex block. The triangle GBF is drawn equal in area to the sum of the triangles GHI and DEF. The whole deformed mass appears, subject to the necessary limitations, to be the triangular block ABC.

that in addition to crustal shortening there may be independent vertical uplift, affecting masses to unknown depths. The history of certain mountain ranges shows that uplift by folding and uplift by other movements have alternated (see p. 238). The calculations also assume that the volume of the rock mass affected has remained constant during deformation. Another difficulty with this method lies in the choice of a reference plane for measuring uplift, and still another arises from the fact that measurement of shortening can seldom be carried far on any one bed, but must be continued on beds stratigraphically above or below which might have behaved differently in the folding. Lastly, the method is strictly two-dimensional, and does not take into account possible elongation and shortening of the rock mass parallel to the trend of the folds.

CHAPTER IX

LANDSLIDES, CREEP, MINE SUBSIDENCE, AND DEFORMATION DUE TO LOCAL CHANGES IN VOLUME

General Account

Gravity causes frequent movements in surface materials in restoring the equilibrium disturbed by deformation, by unequal erosion, by artificial excavation, or by the weakening of rocks due to solution or variation in water content. This is evidenced by landslides, creep, and mine subsidence, and other allied phenomena. Both the hard and the unconsolidated rocks are subject to deformation of this kind. Closely related to many of these gravitational adjustments are expansion and contraction of the rocks, due to changing conditions of surface load, of temperature, and extensive mineralogical and chemical alterations to which surface rocks are subjected.

The structures resulting from surficial movements of this kind are folds, thickening and thinning of beds, duplication and elimination of beds, faults, joints, autoclastic rocks, etc., — all of which have been described in previous chapters in their general aspects. In addition there are many changes in topography represented by scarps, landslide beaches, and disturbance of drainage. In describing surficial movements, therefore, we are considering much of the subject matter of the preceding chapters from another angle. It is desirable, however, to call attention to the prevalence of surficial movements and to indicate something of the variety of conditions which produce them. After all, these are almost the only movements which can be directly observed in process, and their interpretation throws light on rock structures which have been developed in the past. It is coming to be recog-

nized that much of the deformation now observed in the older hard rocks is inherited from a period when the rocks were unconsolidated and under surface conditions. Structures formed at the surface, of course, persist after the rocks are hardened and buried. When they are seen in the older rocks it is not easy to distinguish them from tectonic structures formed in hard rocks under subsurface conditions. It is necessary at the present time to recognize the criteria for their separation and to search for new ones.

The close relation between the present-forming surface structures and the structures exhibited in the geologic columns has not been as apparent as it might be in the writings dealing with structural geology. A considerable volume of literature on current surface movements has accumulated which has not been fully incorporated into the routine presentation of the elements of structural geology. With the increasing application of geology to engineering problems, the importance of more adequate consideration of these surface movements is becoming more apparent. Engineers, perhaps, are more concerned with what is happening now, and what is likely to happen in the near future, than with the interpretation of old structures, but they cannot afford to ignore the evidence furnished from the older rocks.

Slides Due to Natural Causes

Hundreds of landslides, large and small, have been recorded in the last few decades, and probably there are many more unrecorded. It would seem that the earth is seldom free from movements of this sort for any considerable time. If they were as frequent in the past as in recent times the numbers of landslides which have dislocated the earth's materials in the geologic past must have run into millions.

The slides affect both hard and soft rocks, both under air and under water.

Some of the largest slides have moved masses of rock
measured by square miles in area, and several hundreds of fee
in thickness. There are many more smaller ones which are no
sufficiently noticeable to attract attention. In fact the proces
of sliding *en masse* grades into the individual movement o
small fragments, — which is not ordinarily considered as
phenomenon of landslides, but one of ordinary erosional trans
portation. It is clear also that ordinary transfer by erosio
may in time accomplish a mass movement similar to the result
of a landslide, though usually to be distinguished by nev
arrangement of materials.

It is difficult to say what the limit of size may be. It i
altogether likely that there are many important movements o
small displacements which affect so large an area and so gradu
ally that they are not easily apparent. These can be checke
only by careful geodetic observations, which are not yet suffi
ciently numerous to afford any very conclusive testimony, —
except to the fact that the large-scale movements are probabl
taking place. It is even supposed, on the general ground of th
weakness of rocks, that considerable sections of continent
may creep toward ocean basins.

Landslides may be catastrophic in nature, or they may pro
ceed very slowly with a creeping motion, or the movemen
may be an irregular and intermittent one. Even the mos
sudden movements are likely to be measured in hours or days

The angle of slipping may be exceedingly low where ther
are favorable conditions for slipping and strata are weak. I
other cases it may be nearly vertical. A very common cas
is a curved plane of slipping, steeper above and flattenin
below.

Landslides seem to be caused by an immense variety o
conditions. The mass may be so weak that it may fail t
support itself even at a low angle of repose, like viscou
liquid. The slipping of clay and mud deposits and talu
slopes are cases in point. The existence of a plane of slip
ping under a loaded area favors slides. The slip plane ma

be caused by inherent weakness of certain layers, such as mud or clay; by additional weakness caused by saturation with water, which may in effect lubricate a certain zone; by the drying of certain sediments, or by bedding planes, joints, and cleavage. Rocks are constantly undergoing a wide range of chemical and metamorphic changes, all of which affect their strength. The surface contacts of rocks of widely differing strength, such as granite and shale, sometimes serve to localize movements.

Sliding is ultimately the effect of gravity, but more immediate causes are likely to serve as the trigger to set the movement going. Such for instance is the great accumulation of sediments or volcanic products at some one place, increasing the load beyond the resisting power of the rocks and structures beneath. Or, with or without increase of load, the movement may be set going by the erosion of competent key beds on lower slopes, thereby removing the support for masses lying up the dip. Rock masses strong enough to support themselves against the steady pressure of gravity may be incompetent to withstand an earthquake shock. Unusual steepening of slopes from any cause may localize landslides. Young block mountains with steep fault escarpments are characterized by numerous landslides. Glaciers may cause movement by their loading effect, by the thrust they exert, or by sapping or oversteepening the rocks at their source, causing cirques. From this cause in 1920 about 5,000,000 cubic yards on Mt. Blanc moved 8 kilometers. In detail the conditions about glaciers which lead to minor deformation in glacial and underlying materials may be very complex.

At the point of break in landslides, on the up-slope side, fissures are developed which often widen rapidly as the mass slides. Some of these fissures would be known geologically as joints, others as faults. In most cases the fissures, which are multiple, seem to be tensional; they are often, though not always, accompanied by a slight tilting of the sliding mass away from the solid bank. As the surface of slipping is fol-

lowed down, it seems in most cases to flatten and the movement along it is a shear. In addition to the fractures, drag folds, thickening and thinning of strata, duplication and elimination of beds, have all been noted. At the toe of the landslide, movement may be locally upward due to the pressure from behind. Here the movement may be of the nature of a thrust fault or overfold, with the overhanging side going forward.

With these preliminary generalizations, a few typical landslides and subaqueous slips may be described. The term landslides is used in a general sense to include not only the mass movements but the various folds and faults accompanying them.

In 1903 about one hundred acres of clay on the bank of the Lievre River, Quebec,[1] slipped across the river for a distance of four hundred feet and was piled up in masses from 20 to 30 feet high on the opposite bank. The mass moved forward partly as a block but generally in a much broken-up condition. The movement was a pivotal one. Numerous heavy fractures in the mass have a general course at right angles to the direction of movement. In places huge masses of the clay were forced upward along these fissures and show striated and smoothed surfaces. It is thought that the cause of this disturbance was loading by saturation of the clay beds, which are arenaceous in places, and the softening of some interstratified silty layer which was apparently about twenty feet from the surface.

The Columbia lava beds in the Cascade Mountains of Washington rest in places on clays and sands or on deposits of volcanic lapilli. The erosional undercutting of these soft materials furnishes conditions favorable for landslides of the lava beds in the steep escarpments above, and many such movements have occurred. The presence of many vertical joints in the flat-lying layers of basalt furnishes conditions for

[1] Ells, R. W., The recent landslide on the Lievre River, P. Q.: Ann. Rept., Can. Geol. Survey, vol. 15, pt. AA, 1904, p. 136.

the initial breaks. According to Russell[1] the blocks thus broken are tilted toward the up-slope during their downward movement.

In 1903, a mass of rock about one-half mile square and from 400 to 500 feet thick broke away from Turtle Mountain, near Frank, Alberta, slipped to the base of the mountain, across the valley, and 400 feet up the other side of the valley, covering about one square mile to a depth of from five to 150 feet. The rocks were limestones above, dipping toward the mountain, and weaker sandstones and shales below. They were cut by numerous fracture and joint planes. It is supposed

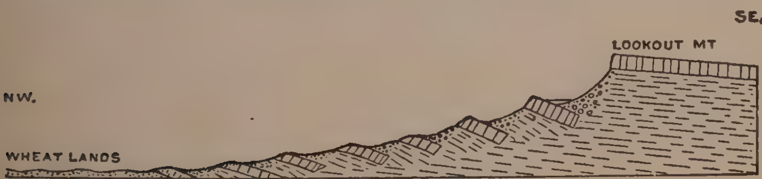

FIG. 81. Ideal profile of landslides on the northern side of Lookout Mountain, Wash. After Russell.

that this slide was due to the steep face of the mountain with its weak base of shale and sandstone, and that contributing causes were probably earthquake tremors in 1901, a preceding period of heavy precipitation and heavy frost, and the mining of a coal seam along the foot of the mountain.

In 1920 vast areas of loess in the province of Kansu, China, slipped during a heavy earthquake. The slides tore away terraced hills, burying or carrying away villages, damming stream beds, turning valleys into lakes, and otherwise greatly changing the topography. Some of the scooped-out places were half a mile in width at the mouth, extended back into the hills for a mile, and furnished enough dirt to cover several square miles of valley floor. " Some were as regular as if they had been made with a gigantic trowel, while others were as

[1] Russell, I. C., Cascades of Northern Washington: 20th Ann. Rept., U. S. Geol. Survey, pt. 2, 1900, p. 194.

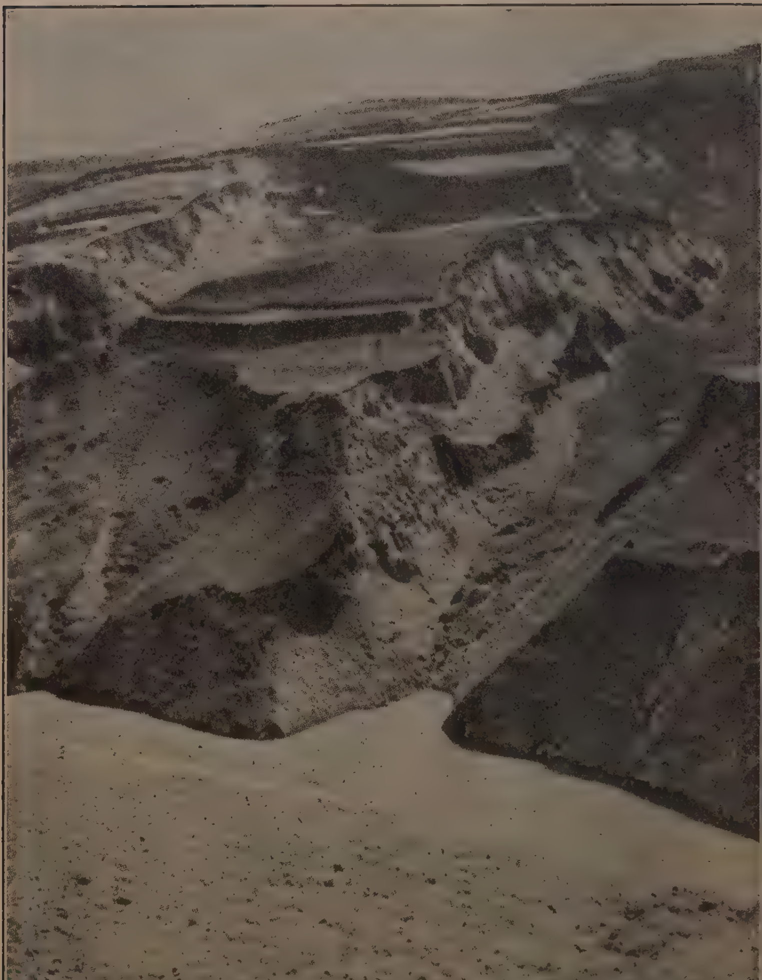

FIG. 82. Landslide in loess deposits of China. All that is left of the terraced field in the middle distance is the little island of earth. Apparently the island is firmer than the earth on either side of it, for it divided the avalanche of dirt into two mouths. (Nat. Geog. Mag., May, 1922.)

ragged as if they had been ripped out of the hills by the teeth of some monster."[1] Not only were many escarpments, fissures, and faults formed, but the earth was thoroughly churned and cascaded like water, forming vortices, swirls, and all the convolutions into which a torrent might shape itself.

Soft muds, clays, and marls in process of deposition under water, and particularly in deltas, slip under load, and develop many breccias, crinkles, contortions and slips. A common feature of these occurrences is their confinement to single beds or zones, between beds which are free from such structures. Grabau[2] illustrates a number of these occurrences in Miocene, Jurassic, Triassic, Devonian and Ordovician beds, of Europe and the United States. These beds have not suffered later tectonic disturbances sufficient to explain these structures. In themselves the structures lack diagnostic criteria which would distinguish them from the results of later tectonic disturbances, unless it be in the general absence of slaty cleavage.

Highly folded and broken limestone layers between undisturbed beds occur at Trenton Falls, New York.[3] In the past there has been some tendency to ascribe these structures to local differences in crystallization or to tectonic disturbances in the hard rocks. The more recent tendency has been to ascribe them to slipping while the beds were soft, perhaps during the process of deposition, more or less in the manner described below by Kindle.

There are other similar structures in limestone, salt, and gypsum deposits to be ascribed to solution, crystallization, coral building, etc., as noted on pages 160 and 224.

[1] Close, Upton, and McCormick, Elsie, Where the mountains walked: Nat. Geog. Magazine, vol. 61, No. 5, May, 1922, p. 463.

[2] Grabau, A. W., Principles of stratigraphy: A. G. Seiler and Co., 1919, pp. 781–785.

[3] Miller, W. J., Highly folded between non-folded strata at Trenton Falls, New York. Jour. Geol., vol. 16, 1908, p. 428; Geology of Remsen Quadrangle: Bull. 126 New York State Museum, 1909, pp. 29–33.

Kindle[1] reports crumpling and contortion of beds, without disturbance of adjacent strata, in the soft sediments of the Bay of Fundy and in the recently deposited sediments of the Avon River, where heavy sand beds are laid down over very mobile mud, — resulting in a squeezing of the softer beds toward the unsupported edges, and the development of a variety of convolutions and faults. He also produces these structures experimentally by differential loading of soft sedi-

FIG. 83. Contemporaneous folding in stratified clay, Courville township, Quebec. After Wilson.

ments. He suggests that the mud lumps of the Mississippi River are merely squeezed-up soft layers due to this process. One of the essential features brought out clearly by the conditions in the Bay of Fundy is the constant shifting, under heavy tides, from sedimentation to scour. Some of the exposures exhibit horizontal beds which have been cut into hollows, which later were filled by sediments in which the laminae partake of the slope of the hollows which they fill and

[1] Kindle, E. M., Deformation of unconsolidated beds in Nova Scotia and Southern Ontario: Bull. Geol. Soc. of Am., vol. 28, 1917, pp. 323–334.

are thus deflected from a horizontal attitude. These are beveled. Horizontal strata next deposited above these irregular surfaces cause disconformity.

During the high-water periods in river beds deep channels with steep walls are often carved in the soft sediments. At low-water stages the escarpments so formed are subject to slump and landslides.

The local disturbances and irregularities developed by original deposition are of especial interest in the study of oil structures. It is clear that strikes and dips taken at such places cannot be used as a basis for interpretation of major

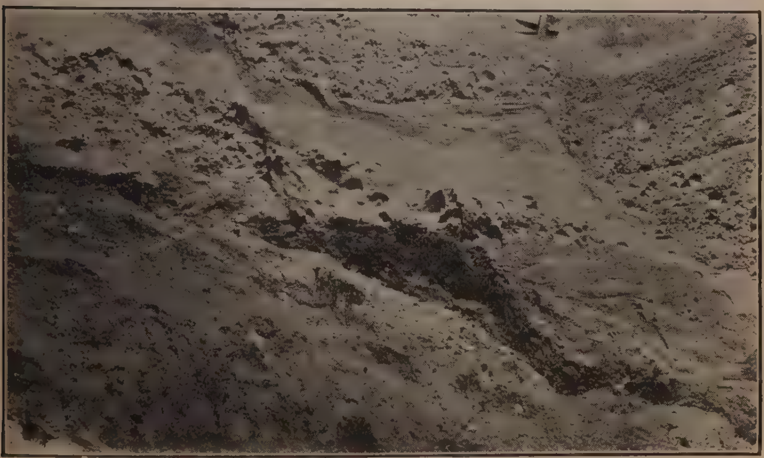

FIG. 84. Faulting in glacial delta deposit near Madison, Wisconsin, due to melting of ice below. After Thwaites.

structures below. The problem is still further complicated by the fact that there are often other irregularities to be explained, due to later solution and deposition, settling, slump, creep of beds, action of plants and frost, etc.[1]

[1] Mather, Kirtley F., Superficial dip of marine limestone strata. A factor in petroleum geology: Econ. Geol., vol. 13, 1918, pp. 198–206.
Shaw, Eugene Wesley, Anomalous dips: Econ. Geol., vol. 13, 1918, pp. 598–610.

Glacial lake clays often show irregularities indicating sliding or drag of grounding icebergs. According to Sayles: [1]

"The ice in coming to rest drags over the bottom, destroys some of the upper layers and contorts many more lower down. Contorted zones produced in this way are the most common irregularities in otherwise regularly banded clays. Glaciated rock fragments are very often found mixed in such contorted zones and a majority of such fragments are found on the top of contorted

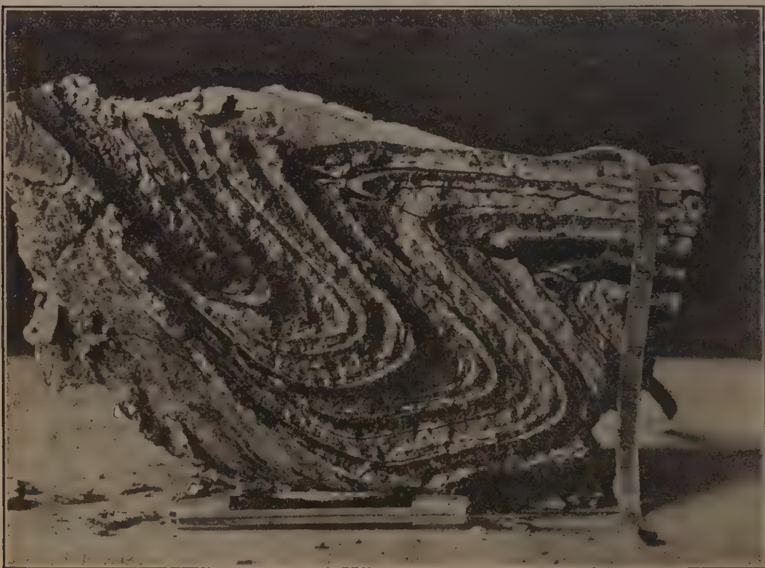

FIG. 85. Contorted bands in glacial clay. After Sayles.

layers, sometimes in a till deposit. In some cases the contorted zones show evidence of actual ice advance. The contortions may be completely cut off on top and the contorted zone itself may have folds eight or ten feet or more from the tops to the bottoms of the arches, with a thin layer of till at the top.

The deformation of layers, intercalated between horizontal layers of the clays, may be due in many cases to shearing produced by

[1] Sayles, Robert W., Seasonal deposition in aqueo-glacial sediments: Memoirs of the Museum of Comparative Zoölogy, Harvard College, vol. 47, No. 1, 1919, pp. 37–38.

ice, well above the deformed zone. The folding observed might result from the downward pressure of the ice alone, or by the dragging effects of moving ice. Inasmuch as the directions of the folds correspond in nearly every case observed in the Connecticut River clays to the direction of ice movement, I have considered it more probable that the results noted were due to ice drag than to vertical pressure alone.

It is also possible that the deformed layers in question might be the result of creep toward an unsupported edge, due to the weight of superincumbent clays alone."

FIG. 86. Crumpled Aftonian (?) inter-glacial deposits distorted probably by being overridden by Kansan glacier, Crow Creek, east of Smithville, Mo. After Hinds and Greene.

The direct thrust of the continental ice sheet has caused many overfolds and overthrusts in glacial materials. An excavation on the campus of the University of Wisconsin disclosed a nearly horizontal overfold or drag-fold thirty feet long, in a position corresponding to the known movement of the ice from the northeast.

Deformation of glacio-fluvial and glacial materials by slumping is observed in many localities. In modern glacial deposits mud avalanches are often seen; these doubtless result in contortion and faulting of the materials. Slumping as a re-

sult of the melting of supporting ice is very widely observed in kames, eskers, pitted outwash, and deltas. In scores of sand and gravel pits throughout the glaciated area, small normal

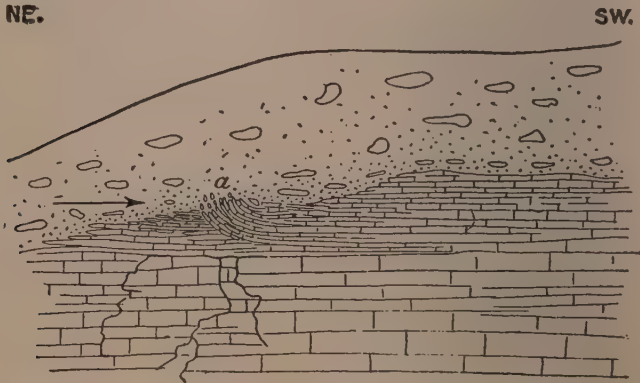

FIG. 87. Effect of glacial ice push on thin-bedded limestone in small quarry northeast of Evansville, Wis. After Alden.

faults and complex minor foldings have been recorded which are due to this cause. Throughout Iowa and Nebraska the Kansas drift contains masses of sand and gravel with disturbed

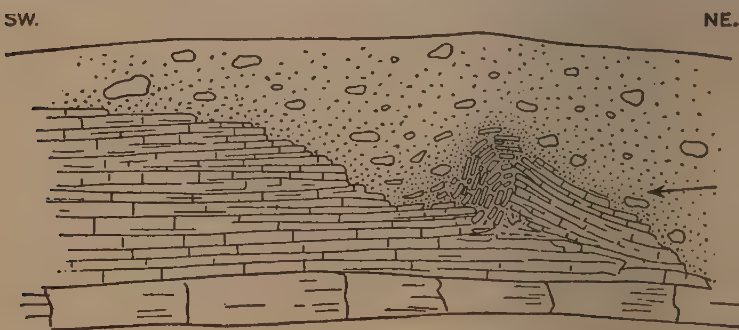

FIG. 88. Effect of glacial ice push; thin upturned beds of limestone in small quarry south of Oregon, Wis. After Alden.

layers; it is probable that these masses were frozen outwash deposits plowed up by the advancing ice.

Glacial cirques contain great masses of rock débris due to

the slipping of the oversteepened slopes. These masses may have the form of small glaciers, and seem to have moved in much the same way.

Expansion of ice on lakes causes local disturbances along the margins, known as ice ramparts,[1] which include folds,

Fig. 89. Ice rampart caused by thrust of ice from lake side. After Buckley.

joints, and faults, Most of these are of a temporary nature, being destroyed in the spring and summer by settling and wave wash.

Slides Due to Artificial Excavations

All of the conditions above described as due to natural causes may be duplicated in slides caused by artificial excavations,

[1] Buckley, E. R., Ice ramparts: Trans. Wis. Acad. Sci., Arts and Letters, vol. 13, pt. 1, 1901, pp. 141–162.

such as railway cuts, mine openings, canals, etc. These constitute an important problem in the engineering application of geology, and there are many reports which throw light on the development of geologic structures. Deep underground mining especially has caused much disturbance in overlying rocks which has been followed out in detail by engineers. In no other cases perhaps have the time and stress elements been so closely recorded and analyzed. This is a vast field which cannot be covered adequately in a book of this scope, and we may content ourselves with a few illustrations which are chosen to show some of the more common phenomena.

The Panama Canal Slides. — The Panama Canal slides[1] occur mainly in a weathered tuffaceous rock, known as the Cucaracha formation, which is weak because of its original texture, because it has been softened by exten-

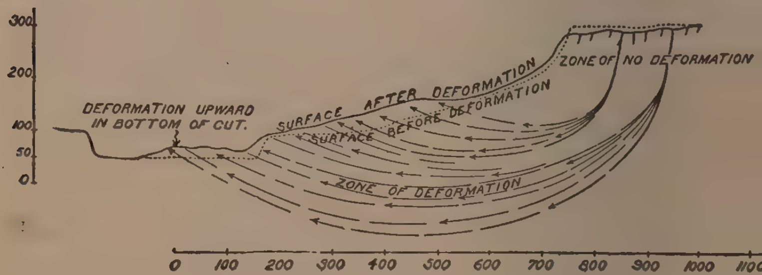

FIG. 90. Ideal cross section to illustrate canalward deformation movements, Panama. After MacDonald.

sive weathering, yielding a considerable amount of colloids, and still further because of local jointing and faulting. Two kinds of slides are discriminated — (1) surficial slips on bedding, joint, or fault planes, or other slippery surfaces, affecting comparatively small masses, and (2) deeper-seated slips along surfaces which steepen above and flat-

[1] Becker, George F., Mechanics of the Panama Canal slides: Prof. Paper 98, U. S. Geol. Survey, 1917, pp. 253–261.
MacDonald, Donald F., Some engineering problems of the Panama Canal in their relation to geology and topography: Bull. 86, U. S. Bur. of Mines, 1915.
Report of Committee of Nat. Acad. of Sci. on Panama Canal slides.

ten below and affect much larger masses. The surficial slips are much affected by seasonal changes in rainfall. During the dry season the surface for a few feet is thoroughly dried out, and when the rain comes it slakes very rapidly and easily washes away. These surficial slips are obviously the result of fracture, mainly of a tensional type. The more important, deeper slips are mainly the result of shear by gravity. There is grinding along multiple shear planes and more or less interior movement of the entire mass, suggestive in general of

FIG. 91. Panama slide of the "fault zone" type. Office stands on ridge formed by basalt dike. The view is northward; a large fault plane, almost parallel to the dike on the north side of it, shattered the rock, which slid along the zone of crushing. After MacDonald.

rock flowage. A large area of ground descends almost vertically and its upper surface gradually assumes a marked tilt away from the canal. The surfaces of the slides are very rough in consequence of the surface breaking in their upper parts. The movement below is more or less like that of a glacier, which flows below slowly and continuously like a viscous liquid, but at the surface is torn and broken into blocks.

"The first indication of a slide of this class is the opening of one or more nearly vertical cracks in the ground at some distance from the Canal. They widen slowly at first and soon may seem to have become inactive; but movement begins again later. It may be months, or even years, after the first appearance of the cracks before important movements of the ground take place. Then the ground immediately in front of the cracks sinks along a slightly curved, but almost vertical surface; as we approach the Canal, the movement of the sliding ground becomes less and less vertical and more and more horizontal. In many cases the bed of the Canal has been forced up vertically, making a high mound, or, after the water was admitted, making an island and blocking up the channel. Evidently, the cracks, the subsidence, the lateral movement and the upheaval, are all correlative phenomena." [1]

The formation is so fine-grained that the rock is kept thoroughly saturated by capillary action, and no real drainage is possible. Therefore, there is no temporary weakening of the formation due to influx of water. So far as water has any effect at all it is due to the fact that it fills the larger cracks and other openings, thereby increasing the load. The committee of geologists from the National Academy of Sciences which has studied these slides concludes that there is little to be gained by coating the surface with an impervious layer in the attempt to keep out water, because capillarity keeps the formation saturated anyway; that the slides start in the dry as well as in the wet seasons; that, however, surface drainage may be a minor help in reducing the amount of water in the larger cracks and openings; that some slight relief might be obtained by unloading the surface at certain points.

Mine Subsidence. — In open cut mines there are slips in the banks subject to much the same laws as govern natural landslides. Commonly the slips are caused by steepness of banks, weakness of rocks, presence of joints, faults, bedding planes or weak layers, affording mobile zones of slipping, and access of water. In unconsolidated material, like glacial drift, vast quantities of water are soaked up which have an effect in

[1] Report of Committee of National Academy of Sciences.

loosening the banks, both because of the load created and because of the softening of certain clayey layers. In general these slides are the result of shear along surfaces steep above and flattening below, with local tensional effects at the surface due to slight tilting of the moving slices. In the great open pits of the Mesabi district of Minnesota great care is taken with drainage, not only to divert the surface water before it enters the banks, but in the drainage of the banks. The ever-present possibility of slips requires the closest study and analysis of the local conditions, and probably more detailed knowledge of the character of landslides has been gained in the progress of mining operations than in any other field.

Of especial interest is the subsidence of ground over underground mining operations, and especially very deep ones.[1] Rocks are almost never strong enough to stand up for any great length of time above a large mine opening; underground mines are usually failing structures. Mining is often conducted on the principle that the roof will fail, and the problem is to conduct the operation in such a way that the failure can be more or less controlled and that it will not interfere with operations. There are a variety of caving methods of mining which are built up on the principle that subsidence shall more or less keep pace with excavation below. Where a large volume of ore is to be removed, with considerable vertical and horizontal dimensions, mining is begun near the top and the overlying cover is allowed to subside more or less gradually as the mining progresses downward. If movement of the overlying material becomes arrested for any reason artificial means are taken to start it, by blasting the roof before the openings beneath it become too large. If mining is carried on too long before movement starts and the openings become too extensive, when the roof is finally weakened there is danger of its coming with such a rush as to destroy all of the openings underneath.

[1] An excellent summary of the literature on this subject is given by L. E. Young and H. H. Stoek in Bull. 91, Engineering Experiment Station, Univ. of Ill., 1916, entitled "Subsidence resulting from mining."

Where thin, flat or gently-dipping beds are being mined, such as coal, natural cement, or Clinton iron ores, the openings may be supported for an indefinite period by leaving pillars. However, by this method a considerable amount of mineral material may be left in the pillars, and even then the roofs will in time fall in unless the rocks are of exceptional toughness, like the gneisses in the Adirondack magnetite mines.

To insure larger recovery of the mineral, and therefore more efficient and economical extraction, a long-wall method of extraction is often used, particularly in thin coal seams; this consists in taking out all of the ore along the working face, leaving no pillars, but filling the ground behind with loose, broken rock to support the roof. The loose rock cannot be packed in sufficiently tight to prevent some subsidence in the roof; as the pressure comes upon the broken rock there is more or less compression and subsidence, which gradually works its way up through the overlying cover of rock, along more or less vertical shear planes, until it reaches the surface. It may take years for this slow subsidence to work through to the surface. In older mining districts where this method has been used, particularly in Europe, the problem of damage to buildings and the area of the surface affected have become important economic problems. In parts of the United States also, where this method has been used for a number of years, the damage to the surface is now becoming apparent, as in Northern Illinois, and these conditions have led to intensive study of mine subsidence which has yielded much valuable information on the strength of rocks and their behavior under various conditions of load and opening.

It is found that the maximum disturbance above is somewhat behind the advancing coal face, but that minor disturbances are projected forward beyond the coal face, sometimes being felt for hundreds of feet in advance. The exact nature of these slight advanced movements is not fully understood. They may be minor sympathetic shears, along flatter shear planes, accompanying the major or nearly vertical shears.

It has been suggested also that they may be the result of tension or pull along the beds, caused by the slump of the beds behind. In one case in Northern Illinois, careful surveys of monuments placed in various positions above an advancing

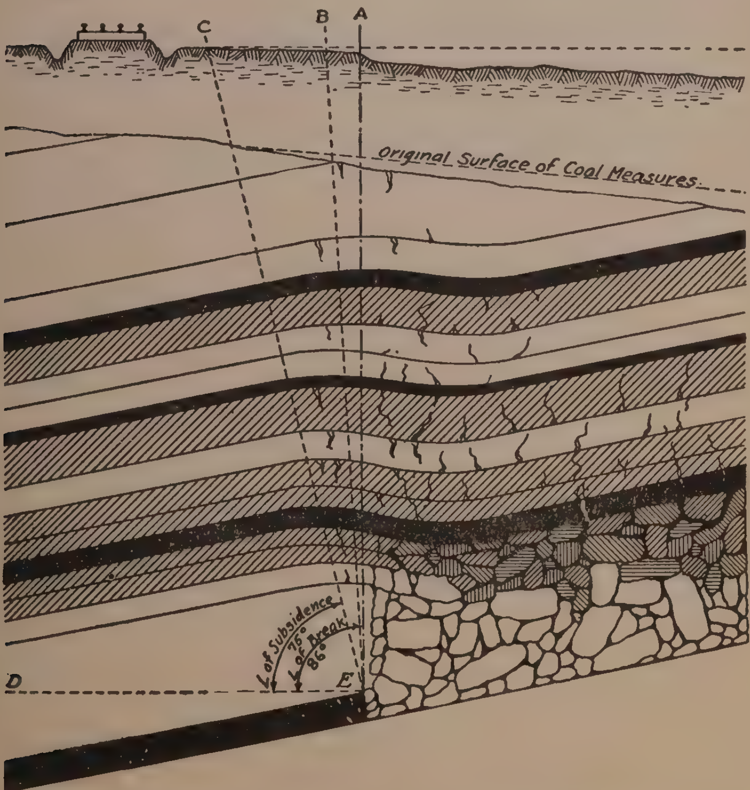

FIG. 92. Subsidence due to removal of coal seam. After Young and Stoek. The lowermost strata have collapsed, the next higher strata have sunk and cracked, and the uppermost strata have sunk without breaking or cracking. There is subsidence beyond the actual angle of break.

coal face seemed to show a preliminary movement of the ground in advance, in a direction back toward the coal face; but the curvature of the beds caused by the subsidence, even

if the rocks had high tensile strength and could carry the entire pull without breaking, would require a movement of only a small fraction of an inch. As the tensile strength of rocks is very small, particularly in large masses where they are cut by joints, it is difficult to prove that disturbance ahead of the advancing coal face is really a result of pull, rather than of shear under gravity.

Where thin beds are standing steeply tilted, and especially where the rocks are of reasonably competent type, openings may be kept open indefinitely by leaving a few pillars, as in the copper mines of Lake Superior. However, even here it is found that when a depth of a mile is reached the pillars are not able to withstand the pressure of the overlying rocks and that they shear off with explosive force, causing small earthquakes. The breaking usually starts with inclined fractures at the corners or salient angles of the pillars.

With the immense variety of conditions in mines, it is difficult to formulate general principles of subsidence which are comprehensive enough to cover all conditions. Factors in the problem are the ultimate strength of rocks, their structure, the number and attitude of zones of weakness, — such as weak beds, joints, faults, and bedding planes, — the conditions of drainage, the depths of openings below the surface, the size and shape of the openings, the rate at which the workings advance, the surface contour, the potential compressive forces existing in the strata containing the workings, the chemical, mineralogical, and volume changes which are constantly affecting rocks near the surface, particularly under the influence of changing drainage.

In general the surface subsidence extends over a greater area than that excavated, indicating that the breaks through which the subsidence is accomplished tend to fan out upward. In gently-dipping strata it has been found that one set of breaks tends to be normal to the bedding. This is a principle worked out by Belgian and French engineers several decades ago, and is known as the "law of the normal." It has especial

reference to gently-dipping coal seams. As the dips become steeper, however, the breaks do not flatten in proportion, but have a more nearly vertical trend than is required by the law of the normal. Also with increasing depth below the ground there seems to be departure from the law of the normal in the same direction.

So far as the mine opening itself is concerned, the first effect of a break may be a so-called arching of the roof. This may not be a true structural arch, but merely an opening left by the fall of rock, which may exist simultaneously with fractures extending far up into the roof. A continued fall of rock

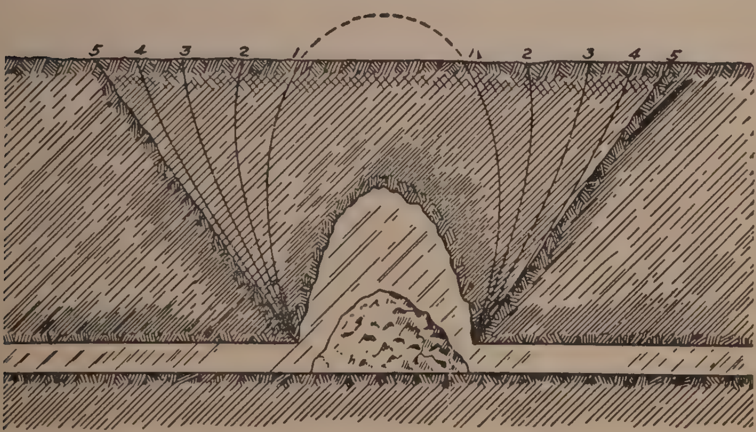

FIG. 93. Subsidence over an undermined area, showing the widening of the subsidence zone above. After Young and Stoek.

naturally extends this arch-like roof fanwise upward until all the ground above falls into the opening.

Subsidence on a large scale is illustrated by one of the Lake Superior mines where a large mass of ore has been removed from a depth of more than 2,000 feet. The strata have a monoclinal dip of about 60°. Along the footwall there are slips or shears, which, near the surface, tend to flatten and cross the bedding obliquely. The upper parts move in the

direction of dip of the shear planes. Tension cracks normal to the principal direction of shear are exhibited in several of the mine buildings. On the upper or hanging-wall side great shear planes dip toward the bedding. They are not exactly normal to the bedding but with a larger vertical element. In general it appears that a wedge-shaped mass is being pulled into the excavated zone. While the width of excavation below is only four or five hundred feet, at the surface the subsidence is affecting an area of several times this width. Nevertheless the *volume* of the openings caused by subsidence at the surface is not as large as the volume of the excavated region filled below, — for the reason that the rock broken by the subsidence has larger volume than in its original state, due to its content of joints, breccias, and irregular openings. Thus it is that a very considerable movement deep below the surface may be much less conspicuous directly at the surface, the subsidence having found expression not only in vertical movements above, but in development of unseen openings below.

It is clear that movements of this sort are essentially shears along inclined surfaces. In the upper parts, however, the subsiding mass is actually pulled away from the solid ground by tension, leaving open crevices.

The subsiding mass, looked at broadly, may move more or less as a unit; but in detail it is much broken, and the several parts move with unequal speed. The distribution of the interior breaks has not been reduced to any general law. About all that can be said is that near the surface the breaks are more numerous and more irregular, and that deeper down more regular shearing planes are apparent. Where the material is soft the movement below may be of the smooth, even nature of rock flowage, though it does not develop the schistose or cleavage structures which are indicative of rock flowage. Striations, mullion structures, and gouges are frequent accompaniments of these movements where the rocks are of appropriate composition and texture.

One of the more important contributions furnished by in-

vestigation of mine subsidence is information as to the time involved. Close checks have been kept of the movements in some mines by placing monuments at the surface and underground, and surveying them at regular intervals, and by plumbing shafts. The dates of appearance of new cracks have been correlated with these records, giving a very good idea of the normal progress of subsidence. It is found that movements are in general slow, being measured in months and years, and that they are not uniform in speed. Somewhat sudden shifts may alternate with periods of quiescence; sometimes also there are catastrophic movements. In the Chapin Mine of the Menominee iron range, the subsidence due to mining 2,000 feet below the surface is so gradual that a railway, built on an embankment crossing the subsiding area, operates without interruption by slowly increasing the height of the embankment to keep pace with the subsidence.

As mining operations become older, more extensive, and deeper, the problem of subsidence becomes more important from engineering, economic, and legal standpoints. The damage caused to overlying mineral seams and to surface improvements and drainage has led to extensive litigation and to large expenditures of money for damages or for correcting conditions. It has been found necessary in Europe to frame laws to meet the problems raised by subsidence. In the United States these problems are just beginning to be serious. Structural geologists are being called in more and more to study these problems, and in the near future there is likely to be considerable increase in our knowledge of their geologic aspects.

Local Disturbances Caused by Change in Volume

In addition to the mass movements of rocks under load, causing landslides and related phenomena, there are surficial movements caused by expansion and contraction of rock masses which may yield all the structures observable in landslides.

These are often difficult to distinguish from the results of mass-slipping. As sediments dry out there is volume decrease and slump, which manifests itself in many joints and faults and perhaps locally in the folding of weaker layers. In the cooling of igneous rocks faulting and jointing are well recognized results of the volume change, and some earthquakes have been attributed to this cause. Successive periods of wetting and drying of sediments, or heating and cooling of rocks may yield most complex results. The weathering of igneous rocks in general causes volume changes, both increase

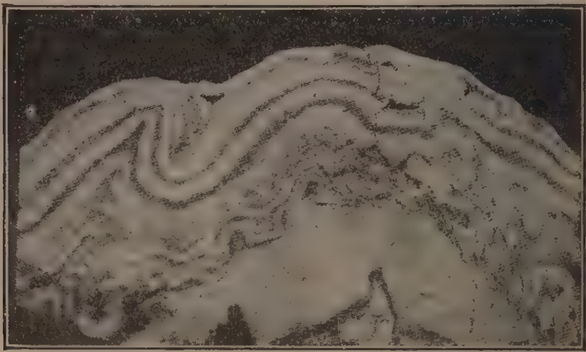

FIG. 94. A specimen of gypsum from Hillsborough, New Brunswick, showing highly folded layers between less folded layers. After Miller.

and decrease, which may manifest themselves in local structural irregularities. The frequent slips and gouges and striations seen in highly weathered rocks, particularly igneous rocks, immediately suggest local volume changes and not any major tectonic disturbance. The swelling of gypsum has long been recognized as a source of local deformation. It is supposed that the primary deposit is mainly anhydrite, and that the hydration to gypsum goes on during deposition and later. The volume increase is 60 per cent. Gypsum layers are characteristically crumpled and contorted in response to the requirements of volume change. Salt deposits exhibit similar phenomena.

The slump of ore-bodies during leaching in the zone of weathering is a well-recognized cause of joints, faults, and folds. One of the best known cases is in the Mesabi iron ore deposits of Minnesota, where the slump has been determined quantitatively, and correlated with known chemical and mineralogical changes. This information is useful in exploration, because when slump is discovered in the layers of the iron formation, concentration of ore can be inferred in the zone below.

In general the structures caused by volume change are local, discontinuous, and unrelated to major tectonic features.

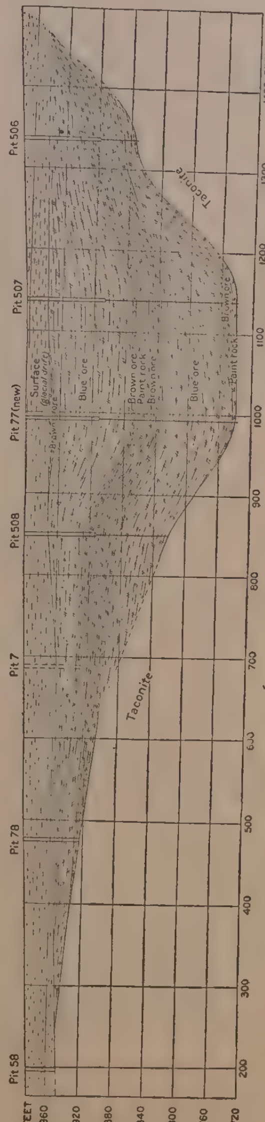

FIG. 95. Cross-section of a Mesabi ore-body showing slump of layers due to removal of silica during concentration of the ore. It will be noted that the ore layers bend down sharply at their contact with the adjacent rock, and that the maximum slump exists over the deepest part of the ore-body. Mon. 52, U. S. Geol. Survey.

Experimental Deformation of Soft Rocks

Experiments on jointing and faulting have naturally been largely confined to hard, brittle materials. Softer materials have been used mainly in the attempt to reproduce folding; in these experiments not only folding, but faulting and brecciation, have been secured in the greatest variety. Willis' well known experiments on the mechanics of Appalachian

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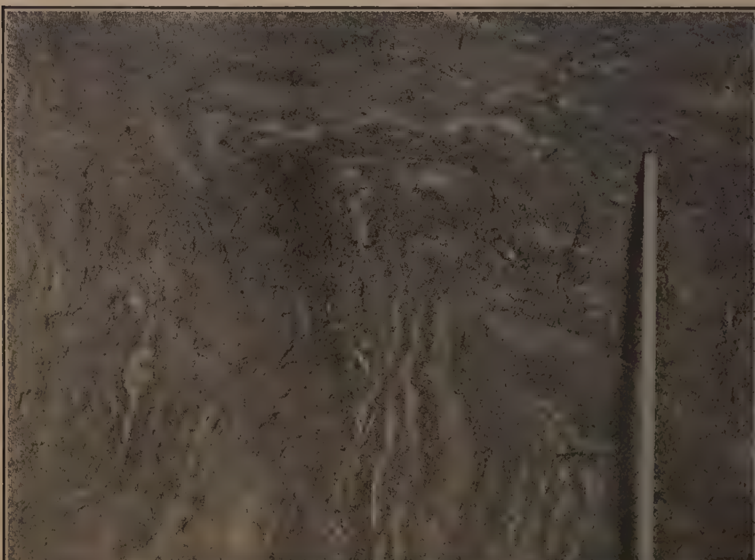

B

FIG. 96. Folding of iron-ore beds, resulting from slump following leaching of silica, Mesabi range. Mon. 52, U. S. Geol. Survey.

structure [1] were with materials more closely approximating the soft-rock than the hard-rock condition. In the structural and sedimentation laboratories of the University of Wisconsin a great variety of structures has been secured by the distortion of clays, sands, and marls. One of the most interesting of these experiments is photographed in Fig. 97, showing soft fragmental sediments, originally slightly tilted to correspond

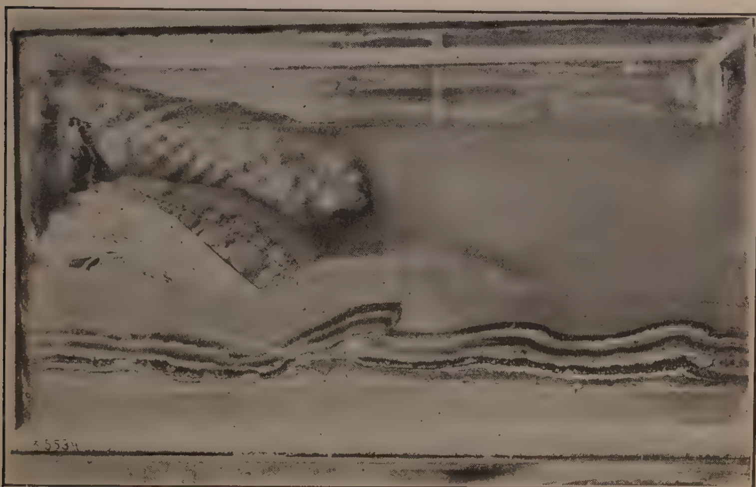

FIG. 97. Experimental deformation of soft sediments by loading. After Rettger.

with initial dip, and loaded above. The softer layers creep down the dip, causing folding and thrust faulting. One of the significant facts brought out by this experiment is that during the deformation there is more or less interpenetration of the materials of the different layers, — suggesting a criterion for the identification of such structures in the field.

[1] Willis, Bailey, The mechanics of Appalachian structure: 13th Ann. Rept., U. S. Geol. Survey, pt. 2, 1893, pp. 211–281.

Criteria for Distinguishing Structures Developed in Unconsolidated or Molten Rocks from Structures Produced in Hard Rocks

Notwithstanding the many observed evidences of movements in unconsolidated and molten rocks, the resulting structures have not been satisfactorily discriminated from the results of hard-rock deformation. The two are usually described and considered independently, and in the study of structures in hard rocks there has not been sufficient consideration of the fact that some of these have been inherited from a soft-rock stage, and are not the result of deformation of hard rocks. It is becoming an increasingly important problem in structural field work to discriminate, not only successive periods of deformation, but the condition of the rocks at the time of deformation. Recognition of the fact that certain structures now existing in hard rocks were formed there when the rocks were in different physical condition entails inferences as to the diastrophic history of quite a different kind from those based on the assumption that the rocks were crystalline and competent at the time of deformation. Some of the criteria noted below are not as decisive as might be desired, but they may serve to indicate approximately our present knowledge of this important subject. They are presented in two groups: (1) criteria for the separation of structures formed in hard rocks from those formed in unconsolidated sediments, and (2) criteria for the separation of structures formed in hard rocks from the structures formed in molten masses.

Criteria for the separation of structures formed in hard rocks from those formed in unconsolidated sediments. (1) Soft rocks are not competent to maintain cavities caused by jointing, faulting, and brecciation to the same extent as hard rocks. Cavities are formed, but they are likely to be small, discontinuous, and local. Joints and faults are likely to be more irregular and ragged.

(2) Soft rocks are not sufficiently competent to transmit

stresses long distances, and hence in general the resulting structures are likely to be more local and discontinuous than in hard rocks. Extensive and uniform joints and faults and joint and fault systems are not the rule. Folds are not likely to be on a large scale, — though to this there are exceptions, as in the cases cited on pages 207–208 of folds caused by the squeezing of soft layers under the vertical load of overlying sediments, or in the case of folds reflecting buried topography, cited on page 160.

(3) The fractures induced in the soft-rock stage are likely to be cemented by the rock material itself, which has either flowed into the openings as a viscous mass or dropped in by sedimentation, either chemical or mechanical. Where the cement is a result of sedimentation, it may show a bedded structure; where it is the result of flowage, it may show a slight parting parallel to the direction of flow, but not a true flow cleavage. Where fractures are formed in hard rocks they either remain open, or are cemented with vein material; or, as in the soft-rock fractures, the cavities are filled by the flowage of softer adjacent layers into the openings, in which case the flowage is likely to be marked by flow cleavage.

(4) There is likely to be less sharp definition of structures produced by soft-rock deformation than in hard-rock deformation. Experiments show that there is a certain amount of interpenetration of the grains of adjacent beds when they are deformed in a soft condition.

(5) So far as the writer knows, a true flow cleavage characterized by the development of secondary platy minerals is confined to hard-rock deformation. Disturbances in the soft stage may develop a rude, platy structure or parting, inclined to bedding, or may accentuate shaly partings, but they do not seem to cause a schistose structure inclined to bedding, built on abundance of platy minerals. Close and intricate folding in a crystalline shale formation, without conspicuous fracture, and without the accompaniment of flow cleavage, suggests deformation in an early, unconsolidated state.

(6) The folds, faults, joints, and breccias, or intraformational conglomerates, and mud cracks formed in soft rocks may be confined to individual layers in the midst of a conformable succession of beds, where it is clear there has been no extensive time interval between the deposition of the underlying and overlying beds, and that the structural disturbance is a mere incident or temporary interruption of an otherwise continuous process of sedimentation. The structures indicate that the mass had become sufficiently coherent to fragment, though perhaps far from the hard-rock state. The hardening and cracking may be the result of temporary exposure to the air, causing mud cracks and more irregular brecciation, or they may be the result of current or scour action under water. Usually these structures are not extensive and do not indicate a movement in a uniform direction over a large area.

(7) The nature of the deposits may furnish presumptive evidence as to the conditions under which the deformation took place. For instance, in delta deposits slips and overfolds are more likely to be assigned to soft-rock deformation than they would be in a uniformly bedded marine sand or quartzite formation.

(8) Local structures which do not furnish satisfactory criteria may be so related to large regional structures as to indicate a common origin. For instance, local drag in beds may be in the direction required by the major displacements, indicating correlative origin. It is often possible to assign the larger structure to orogenic movements later than the consolidation of the rocks, in which case the local structure may be similarly correlated.

(9) The relation of the deformation to the erosion surface, whether a present or buried one, may give a clue as to the date of a structure, and, therefore, to the probable degree of consolidation of the rocks at the time of deformation.

In the foregoing criteria reference is made in several places to the local and restricted scale of the soft-rock structures. As a matter of fact this is not a criterion to be relied on without

confirmatory evidence. Scale is only relative, and in classifying a given structure there is no set standard for comparison. Some hard-rock structures are also local, and some soft-rock structures are extensive, perhaps more so than now recognized.

Criteria for the separation of structures formed in hard rocks from structures formed in molten masses. Movements of rock masses in the molten stage produce gneissic structure, folds, faults, brecciation, and so on, which are sometimes difficult to distinguish from the results of later movements after the rocks have become hard and crystalline. A really adequate consideration of the criteria for their discrimination extends beyond the field of structural geology into that of metamorphism, petrology, and vulcanism, and will not be here attempted. We confine our comments to a few of the more obvious structural features.

Some of the criteria for distinguishing the primary igneous gneisses are summarized by Trueman [1] as follows:

"*Field evidence:* Banding in apophyses from the gneiss parallel to the walls and at an angle to the schistosity of the enclosing rock,[2] dikes of pegmatite belonging to the same magmatic series as the gneiss and either parallel to the gneissic structure and foliated with it or cutting the gneissic structure and undisturbed; lack of sharp contact between the acidic and more basic portions of the gneiss, indicating high temperature during the solidifications of the different bands; [3] presence of inclusions of foreign rock, which are but slightly deformed, in a matrix of well-banded gneiss; [4] presence of distinct bands of widely different composition, none of which may show evidence of shearing; flowlike curves of the banding, some of which may close in a circle."

"*Mineralogical evidence*: Presence of minerals formed characteristically only from igneous melts and arranged in a manner impossible of formation from solid rocks by metamorphism, e. g., nepheline and olivine; textures due to crystallization from an

[1] Trueman, J. D., The value of certain criteria for the determination of the origin of foliated crystalline rocks: Jour. Geol., vol. 20, 1912, p. 231.

[2] Gregory, J. W., Quart. Jour. Geol. Soc. London, 1894, p. 265.

[3] Geol. Survey Can., Mem. 6, 1910, p. 83.

[4] Geol. Mag., new ser., decade 4, vol. 4, 1897, p. 354.

igneous melt. Weinschenk [1] considers that epidote, garnet, clinozoicite, sillimanite, and chlorite crystallize from the magma in the case of primary gneisses on account of the pressure present during the solidification of the rock, but the exact state of the rock during their formation is not definitely known."

Trueman points out that some of the features of mineral elongation described as characteristic of the crystalloblastic structure of metamorphic gneisses are, as a matter of fact, possessed by the primary gneisses, and that the texture of the primary gneisses as a whole appears to be intermediate between the igneous and metamorphic types, being more like the latter, according as the movement producing the banding continued later into the period of consolidation of the rock from the melt.

The use of texture as a criterion for the identification of primary gneisses seems on the whole, then, to be only of limited application, even if it be true that the primary gneisses as a group have textures of igneous rocks while the secondary gneisses have, as a whole, the crystalloblastic textures.

The problem of distinguishing origin of gneisses is even more complicated for the so-called injection gneisses, formed by the intrusion of igneous material in thin leaves between laminae of the country rock. These gneisses may have characteristics of both primary and metamorphic gneisses.

As a whole the banding in primary gneisses lacks continuity, and the folding of the bands is far more irregular, local, and involved than that resulting from later dynamic movements.

Schistosity is less abundant in igneous gneisses than the gneissose structure (see p. 146).

Faulting and brecciation developed during the molten stage are also local, intricate, and less regular in distribution or in indicating any uniform direction of movement, than similar structures developed during the hard-rock stage. The open-

[1] Weinschenk, E., Congres' geol. inter., compte rendu, session VIII, I, 1900, p. 340.

ings are filled with crystallized igneous rocks similar to the broken fragments, in contrast with the frequent vein fillings developed in the hard-rock stage (see p. 229). Rock flowage subsequent to the fracturing of hard rocks may likewise fill the spaces with rock material; but this is likely to be highly schistose, and the angular fragments become more or less rounded and elongated.

CHAPTER X

MAJOR UNITS OF STRUCTURE: GEANTICLINES, GEOSYNCLINES, OCEAN BASINS, CONTINENTS, MOUNTAINS, PLATEAUS, POSITIVE AND NEGATIVE ELEMENTS[1]

We have discussed the evidences of rock deformation which ordinarily come within the range of our direct observation — the units of structure with which field work ordinarily deals. Looking at the earth more broadly we see continents and ocean basins, mountain systems, plateaus and depressed land areas, geanticlines and geosynclines, and other great lineaments, all of which have been produced at least in part, though not entirely, by deformation. A *geanticline* is a gentle upswelling of a wide area, and a *geosyncline* is a gentle downwarping. They may be independent of local structure; a region intricately folded and faulted may warp up or down. Furthermore, a geosyncline may be accentuated by erosion and when later depressed and filled with sediments, it may be difficult to determine how much of the depression is due to downwarping and how much to erosion; yet it is often called a geosyncline of deposition. Some portions of the earth's crust have tended during geological time to rise intermittently and thereby remain uncovered by marine sediments. These have been called *positive elements*.[2] Other parts, called *negative elements*, have been submerged again and again during geologic history. The pre-Cambrian shield of North America

[1] No attempt is made under this heading to present a comprehensive account of the actual morphological features of the earth's surface. A start in the study of this immense field is best made by reading Suess's great classic, "The Face of the Earth," which has served as the foundation of most later discussions.

[2] Willis, Bailey, A theory of continental structure applied to North America: Bull. Geol. Soc. Am., vol. 18, 1907, p. 390.

is a positive element; the Paleozoic area of the Mississippi Valley is a negative element. These divisions are necessarily vague and their boundaries have shifted widely during geologic time.

Erosion and deposition play a large part, often a dominant part, in determining the topography of these larger units. Treatment of their structural features, to which these pages are confined, therefore covers only part of the story. The other part is covered mainly by the science of *physiography*, which deals largely with the description and origin of the features of the earth's face, and by *paleography*, which deals with the sedimentary record and the reconstruction of land and ocean areas in various periods of the geologic past. Structural, physiographic, and paleographic geology are so closely complementary that no one can advance far without using the others, — a fact which has only recently become apparent to investigators in these fields

The units of structure here to be considered are on so large a scale that they are usually regarded as involving deep movements within the earth, — deeper than those causing folding, faulting, and jointing. Vertical uplift or depression may be more conspicuous than evidence of horizontal shortening, though horizontal shortening may be apparent when the structures as a whole are considered. Dominantly vertical movements of this kind are perhaps more often of an epeirogenic than of an orogenic nature. There is abundant paleographic evidence of the recurrence of epeirogenic movements at many times and places in the earth's history. In fact, such movements are seldom absent; the earth's surface seems to be constantly bulging at some points and receding at others.

It may be noted that uplift and depression are relative terms. The lowering of the sea bottom may cause a continent apparently to rise even though it has really stood still. Or both may be sinking but at different rates, — in which case apparent uplift or depression merely expresses the difference

in rate. Actual uplift or depression is very difficult to prove by inductive reasoning from direct observation.

The larger units of structure related to epeirogenic movements may exist independently of, or together with, the smaller units of orogenic origin. Where together their relative distribution may suggest common genesis, though this has been actually proved in but few cases. It has been noted that plateaus and mountains and continents may be bordered on one side or another, or both, by great thrust faults which dip underneath the higher areas, or by folds. The peripheral arrangement of these structures suggests that they are but incidents in the deformation causing the major structures.

These major elements and their genetic relationships to local structures are naturally of great interest to students of earth origin and development, and much effort has been given to correlate them, and to generalize from them as to the behavior of the earth as a whole during its long development.

As one notes evidences here and there of some order and synchronism, he cannot but suspect that the major earth deformation as a whole may ultimately be reduced to simpler terms than a casual inspection of the irregularities of the surface might suggest; but as yet such attempts have not been highly successful, due to the facts that our knowledge of the earth's surface is far from complete, and that the statements of basic facts have often been vitiated by initial assumptions as to the origin of the earth and the causes and distribution of the ultimate forces. Observers have read into the structural complex of the earth certain broad lineaments which they would like to see there. It is possible to start with almost any hypothesis and to find somewhere evidence which is accordant with the initial assumption. This has been done with apparently equal success for fundamentally conflicting hypotheses, and it is almost impossible to be sure that accordance of evidence with one hypothesis is really valid proof of that one to the exclusion of another. In short, our inductive treatment of this problem is in a very elementary

state, and it may be a long while before the basic data are adequate to warrant inductive generalizations free from preconceived notions as to what ought to exist. This may be said without disparaging in the least the efforts of those who have tried to see beneath the surface. It is only by persistent trying that the truth is ultimately made to reveal itself. Our purpose is merely to call attention to the fact that the explanations of the broader structural features of the earth are yet far from verified by direct observation.

Any serious attempt to present the present status of the investigation of the earth's major units of structure would unduly extend this book, involving as it would consideration of various hypotheses of the origin of the earth. While such consideration is a logical consequence of a study of structural geology, so many other sciences and factors are involved that we should soon be far away from our main thesis — observed evidences of rock deformation. We may confine ourselves, therefore, to mere mention of some of the best established facts and hypotheses related to major units of deformation.

Mountains

Types of mountains. Mountains may be carved by erosion from undeformed sediments or undeformed igneous rocks. They may be formed entirely by volcanic accumulations without the aid of erosion or secondary deformation. Most of the larger mountain ranges are sculptured in rocks which have undergone secondary deformation and uplift. They are called *tectonic* mountains. Depending on the nature of the deformation, they are called also fault-block mountains, monoclinal mountains, synclinal mountains, etc. They are commonly dated from the time of deformation and uplift, rather than from the period of erosion. Uplift relative to sea level must precede erosion, and in that sense is primary and essential to mountain building. The uplift, however, may produce a plateau or other forms quite different from mountains. Dif-

ferential erosion therefore is necessary to produce the forms of mountains. In time erosion completely base-levels mountains, as it has so largely in pre-Cambrian areas. In the highest existing mountains the uplift and deformation have been of recent date, and erosion has not had time to reduce them.

Attention is here especially directed to *tectonic* mountains as conspicuous surface expressions of the structures described in the earlier pages of this book. We may indicate only some of the simpler relations between structure and mountain ranges. In many ranges there have been repeated deformations and uplifts and repeated cycles of erosion, which have left the present topography in relations to structure which cannot be simply stated. It has often been possible to work out the complex history of the relations between structure and erosion in the development of the present topography, but this involves also the use of principles of physiography and will not be entered into here.

Mountains and normal faults. Some mountain ranges are the result dominantly of nearly vertical movements along faults. They stand above the adjacent lowlands by reason of more or less vertical faults which bound them. German geologists call them *Horste.* The down-throw blocks between horste are called *Graben.* The Great Basin ranges of the western United States are classic illustrations of this type, though there is a difference of opinion among geologists as to the relative importance of faulting and erosion in producing their present topography. The published discussion of the origin of these mountains illustrates the trend from an earlier emphasis on structural features, such as faults, toward a wider recognition of the importance of erosion (see p. 104). The Hartz Mountains of Germany are another example.

Horizontal fault dislocations are conspicuous in some ranges, as for instance in the Dutch East Indies, where earlier structures of folding are horizontally shifted. Steep-dipping faults

in general are likely to be marked by sharp topographic depressions.

Mountains and thrust faults. Thrust faulting has played an important part in the growth of many mountain ranges, but its influence on present topography is not to be stated in simple terms. Repeated fault slices piled one on top of another tend to form the present elevations, as conspicuously illustrated in the Highlands of Scotland, in the Scandinavian Highlands, in parts of the Alps, and in the southern Appalachians. Erosion, working on the tilted fault slices, tends to form linear ridges more or less parallel to the fault traces; but this tendency is much masked, especially where the fault planes are of low angle and have curving outcrops, due to erosion along transverse vertical planes or softer beds. The ridges tend to be steeper on the side toward which the overthrusts have moved.

Single thrust faults may border the uplifted areas, as along the east front of the Rocky Mountains of Montana and Alberta, dipping toward the uplifted masses. These seem to be most common on the steeper side and dip in the same direction, though not in the same degree, as the axial planes of any associated folds. This relation suggests either overthrust from the side away from the fault, or underthrust from the side on which the fault trace appears.

Thrust faulting in mountain-making has been supposed to be associated with the deformation of a thin shell. Where a thicker mass has been affected thrust faults are less important (see pp. 241–242).

Mountains and folds. Folding is the most common structural feature of mountains. It is commonly assumed to be contemporaneous with uplift, but it may also precede or follow uplift. The Andes and probably the Himalayas seem to have been folded before uplift.[1] An individual mountain ridge or peak may represent a single simple fold, but for the

[1] Reid, Harry Fielding, Isostasy and earth movements: Bull. Geol. Soc. Am., vol. 33, 1922, pp. 320–322.

most part a mountain range or chain includes many parallel and overlapping asymmetric folds, pitching in both directions along the trend of the range, with more or less cross-bending or warping of the axial lines in horizontal plan, and more or less faulting.

Seldom do the actual topographic surfaces correspond to the forms of folds. In young mountains, where erosion has not had time to greatly modify the surface, the outlines of open folds may roughly correspond to the topographic forms. As erosion progresses there is wider and wider departure from this relationship. Erosion follows down the softer layers, or along faults and joints or other structures easy of attack, leaving an upstanding mass that may not at all correspond in its outlines to the form of a fold. It is more likely to be a syncline than an anticline. The axial lines of anticlines seem to be peculiarly susceptible to erosion and are often cut down faster than the synclines between, thus leaving synclinal ridges. This is well illustrated in the Appalachian region. The stumps of mountains throughout the pre-Cambrian are largely of this type. Where an anticline has a core of more resistant igneous rock, as in some of the Front Ranges of Colorado, there may be accordance of original structural outlines with the topography, even after long-continued erosion.

In an area of monoclinal or isoclinal folding the softer beds are eroded and the more resistant beds stand out as linear ridges, with steep sides generally in the direction opposite to the dip. Fairly regular step mountains or step topography may be produced in this fashion.

As the folding becomes closer and more complicated, the relations to topography likewise are more complicated. The great overthrust folds result in the piling up of strata in the same manner as by overthrust faulting, forming ridges, which in general persist in present elevations; but the varying resistance of the rocks to erosion results in wide variations between the original forms of the folds and the present topography.

As in the case of thrust faults, the steep slopes tend to be on the sides away from the thrusts, with gentle slopes toward the thrust.

The drag type of folding is not uncommon, as for instance in the Appalachian mountains. The trace of a bed in plan is a highly crenulated line running diagonally to the general trend of the range, and suggesting shearing movement parallel to the range (see Fig. 98). Examination of some of the United States Geological Survey folios shows local areas in which the drag folding is uniform in direction; but the study of this feature has thus far been so fragmentary that no generalizations are yet possible as to differential movement of considerable parts of mountain ranges.

Mountains and cleavage. There are mountains in which cleavage is not conspicuously developed, but ordinarily there is cleavage in the softer beds, or near any intrusive igneous core. In general the cleavage is more or less steep and trends with the range. It is parallel to the axial planes of the folds and locally may be nearly parallel to the thrust faults. Where the axial planes of the folds are tilted in a uniform direction the cleavage also is likely to be so tilted. There are of course many exceptions to these generalizations. The roots of mountain ranges which have disappeared, like some of those in the pre-Cambrian, are characteristically marked by cleavage, which by its trend and attitude tells us something about the former mountains.

Depth of mountain deformation. Calculation of the depth of folding by the methods discussed in Chapter VIII suggests that mountains of sharp folding, great thrust faulting, and much crustal shortening result from the deformation of a thin shell, which has sheared over the less yielding base beneath it, as illustrated by the Jura, Alps, Scandinavian chain, Scottish Highlands, the Backbone of Brazil, the Appalachians, and perhaps the Rocky Mountains of Alberta. As representatives of the thicker-shell type of mountains, in which vertical movements are more pronounced, and horizontal shortening

by folding and thrust faulting less conspicuous, there are the Colorado Rocky Mountains, the Cascades on the Pacific Coast; the Western Andes, and the Abyssinian Mountains.[1]

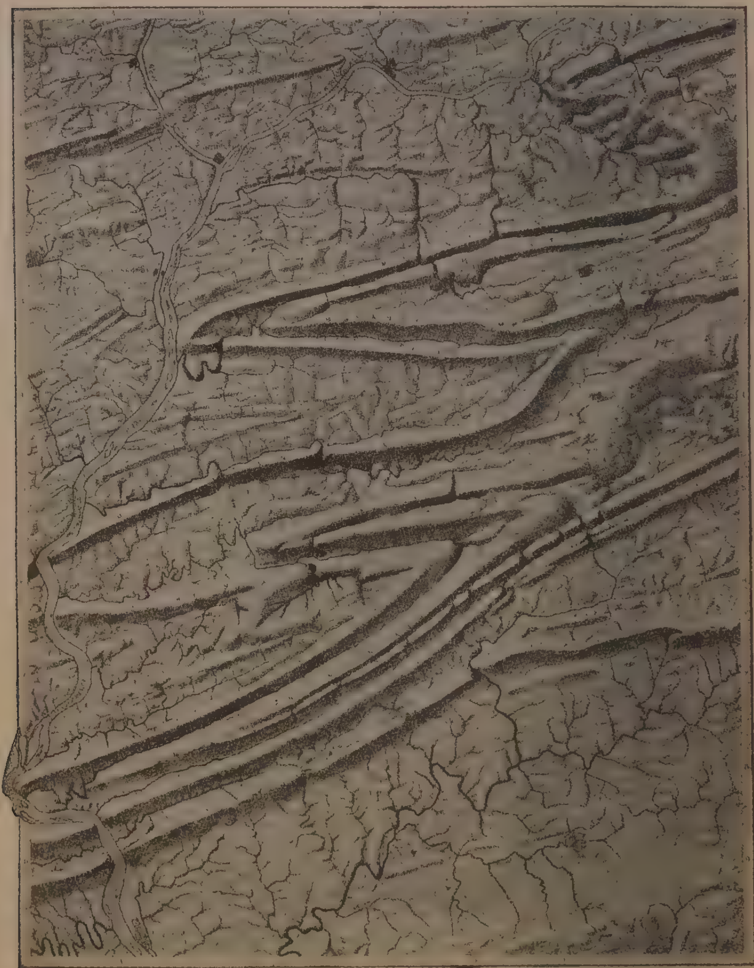

FIG. 98. Relief map of western portion of the anthracite basins, Pennsylvania, showing canoe valleys and mountains and the course of the Susquehanna across them. After Willis.

[1] Chamberlin, Rollin T., The building of the Colorado Rockies: Jour. Geol., vol. 27, 1919, p. 251.

The mass affected in two specially studied cases (see p. 197) seems to be shaped like a downward-pointing wedge, — suggesting that this type of deformation may be a common one, not only in mountains, but even in large elements of structure like plateaus and continents.

The deformation by folding and faulting seen at the surface may be only a superficial expression of deeper and slower movements resulting in rock flowage and folding of quite different distribution.

It should be kept in mind also that folding and uplifting are not necessarily contemporaneous, but that there is evidence in some mountain ranges, such as the Andes and the Himalayas, that uplift has come considerably later. In fact the more that is known about mountains, the more complex their diastrophic history appears. Therefore the depth of the mass involved in the deformation cannot be inferred solely from the surficial shortening.

Crustal shortening involved in mountain-making. It is obvious that mountains caused by folding and thrust faulting have been shortened normal to their trend. There have been various attempts to estimate this shortening by measuring the displacement along thrust planes and especially by estimating the original dimensions of the beds which are now folded. Chamberlin [1] measured a section across the Colorado Rockies and found that 140 miles of country had been shortened into 132 miles.[2] A section across the Pennsylvania Appalachians shows that 81 miles were compressed into 66 miles. Heim [3] estimated that the Alpine area before compression may have been 1200 kilometers across and is now only 150 kilometers. LeConte [4] estimated the transverse shortening of the Coast ranges in California to be from 9 to 12 miles.

[1] *Loc. cit.*, p. 234.

[2] Shepard reports shortening of 37% in the eastern part of the Colorado Rockies. Shepard, F. P., Indications of important horizontal compression in the Colorado Rockies: Am. Jour. Sci., vol. 5, 1923, pp. 403–408.

[3] Heim, Alb., Mechanismus der Gebirgsbildung, vol. 2, 1878, pp. 210–215.

[4] LeConte, Joseph, Elements of Geology, 5th ed., p. 266.

It is safe to say that the shortening involved in mountain-making of this type is very large, but there are so many difficulties in measurement that quantitative accuracy has hardly been reached. The shortening by folding may be partially compensated by jointing, faulting, and intrusion, which tend to lengthen the beds. It is seldom possible to make the measurement on any single bed, because it is not exposed throughout the section. Beds differ very much in degree of folding. In continuing measurement on an off-set bed either more or less shortening may be indicated than in the original bed. Folds may die out in one direction or another from a given bed, so the shortening proved for one general horizon may not indicate the shortening for another horizon above or below. Some folds are merely the result of differential slipping of an upper bed over a lower; the upper and lower beds may not be shortened in the process. Some beds are thickened and thinned to a marked degree by folding, and data are not available to correct for this factor.

Transverse shortening of mountains is not necessarily synonymous with transverse compression by forces acting normal to the range. The pressure may have been rotational or shearing (see p. 190). The best experimental reproduction of repeated folds of the Appalachian type has been accomplished by application of inclined shearing stresses.

Mountains formed by normal faulting of the Basin Range type may involve no transverse shortening; even extension may be possible.

We ordinarily consider only the transverse shortening. Mountain structures, such as the pitch of folds, may show longitudinal shortening. When we consider the great length of mountain ranges as compared to their width, it appears that even if the percentage of longitudinal shortening is less than that of transverse shortening, in the aggregate it may be just as great or greater. However, the fact that folds are pitching does not necessarily mean cross-folding or longitudinal shortening. In an excellent experimental reproduction

of repeated folds of the Appalachian type, Mead[1] showed that there was actual longitudinal extension, expressed by cracks across the axes of the folds.

An accurate measurement of crustal shortening by mountain-making is much to be desired because of its bearing on the problem of earth genesis. Various guesses of the amount of shortening have been used as evidence for and against the several hypotheses of the causes of earth failure (Chapter XV).

Localization of tectonic mountains. Tectonic mountains are localized in comparatively few curving linear zones, more or less branching, cross-bent, and locally interrupted, which may extend over considerable parts of the earth's circumference.

The older mountain chains, particularly the pre-Cambrian, have been eroded until only their stumps remain, and these have been so largely covered by later sediments that their distribution is only partially known.[2] Also later tectonic movements are superposed upon them, and have in some places so twisted the earlier structures that it is hard to make out their original trends. The tracing out of these old mountain ranges remains largely to the future; it presents an attractive field to the student of orogeny. Notwithstanding the scantiness of our knowledge, we seem to recognize a tendency for these old mountain ranges to accord somewhat in distribution and trend with the later ranges. The zones of weakness represented by the earlier mountains seem to have been followed by later mountain-making stresses.

The fact that parts of the older terranes are so complexly deformed has been made a basis for the conclusion that mountain-making processes were formerly exceptionally active and widespread. The Archean period, especially, has been regarded as one of exceptionally active orogeny. In our present

[1] Mead, W. J., Notes on the mechanics of geologic structures: Jour. Geol., vol. 28, 1920, pp. 505–523.

[2] Ruedemann, Rudolph, The existence and configuration of pre-Cambrian continents: 17th Rept. of Director of N. Y. State Museum for 1920–1921. 1922, pp. 65–152.

state of knowledge it is hardly safe to accept this conclusion, for the reason that rocks of these old periods have suffered not only contemporaneous deformation, but all the deformation of later periods, and we are not yet in a position to say positively just how much should be assigned to one period and how much to another.

After the pre-Cambrian the next important mountain-making period is at the close of the Paleozoic. The Appalachian Range of eastern North America dates from this time.

Then came the mountain-building at the close of the Mesozoic and succeeding Tertiary period, represented by the great Himalaya mountain belt of western Asia, and its extension westward into southern Europe and northern Africa, where it connects with the Caucasus, Carpathian, Apennine, and Atlas Ranges. To the northeast the Himalayas connect, through the Aleutian Islands, with the Cordilleran Ranges of the western hemisphere, the extension of which may be represented also in Antarctica. To the southeast the Himalayas branch, and line up with the mountains of the Dutch East Indies. The Pacific is pretty well bordered by ranges of this period. These are the greatest of the mountain ranges and the most recent. They are great because they are recent, erosion not having had time to reduce them as it has the older mountain chains.

A survey of the general distribution of these principal mountain chains brings out the fact that they roughly follow continental margins, and that some of the greatest mountain chains stand opposite the greatest depths. Where they depart from this position evidence has been found in some cases that they were on the continental margins at the time they were formed, even though they are now some distance inland.

In a general way it may be noted also that the mountain lines are looped in a series of intersecting arcs or festoons, with their convex sides toward the oceans, either of the present or of the past. This arc-like arrangement has been regarded by some investigators as the dominant feature of mountain

distribution, and inferences have been built up on this assumption. Mountain chains, like other lineaments of the earth's face, are so immensely varied in their general form and detail, and are the net result of so many different processes and episodes, that their reduction to any simple pattern must, in the present state of knowledge, be regarded with some reservation. The natural desire of the scientific mind to discern broad underlying principles may lead in this case to acceptance of simplicity of pattern which only partly expresses the true situation, and may not take into account other patterns which really exist.

Perhaps the most significant generalization is that mountains have formed in great geosynclinal areas of sedimentation along shore lines, and in periods closely following the deposition. The sedimentary deposits in folded chains are often thick and of a shore type. When traced away from the mountains they are found to be thinner. This is merely an empirical fact of observation, which is not yet established for all mountain ranges. Any attempt to explain why mountain-making should follow geosynclinal deposition is on less certain ground.

In general, geosynclinal areas, lying as they frequently do between continents and oceans, are in places where any adjustments between these major segments of the earth's crust are likely to be concentrated. As will be seen on another page, movements between these major segments seem to be a necessary consequence of almost any hypothesis of earth deformation. Sediments in geosynclines are likely to have a considerable initial dip and when tangential forces later affect them they find it easier to deform rocks already slightly tilted than those lying flat. Still further, the sediments in a geosyncline are usually poorly consolidated, and therefore inherently incompetent.

Another cause often cited for the geosynclinal location of mountains is the heating of the base of the sinking mass of sediments, softening, weakening, or expanding them, and

thereby localizing deformation by general earth stresses, whatever their origin. The mechanism of this process is not entirely clear. It has been suggested also that the geosyncline merely sinks into a softer or molten substratum, and thereby becomes folded.[1] Concurrent with the sinking the liquid rock might rise beneath adjacent parts with lighter loads. Batholiths of granite surrounded by synclinal meshes of schistose sediments in the Canadian pre-Cambrian have been cited as illustrating this kind of movement.

The relation of mountain-making to geosynclines of deposition has perhaps received most attention in connection with the study of isostasy. The loading of heavy geosynclines is supposed to disturb isostatic equilibrium and to start a sequence of movements dependent on volume changes which result in mountain-making (see pp. 332–340).

In general it may be said that the real significance of the relation of mountains to geosynclines of deposition is not yet understood. None of the various explanations offered seems to be entirely adequate. The writer's preference is for the view that geosynclines are mainly zones of structural weakness which are vulnerable to the attack of great earth forces, whatever their origin. It is to be remembered that some mountain deformations are not associated with geosynclines of deposition, that there are geosynclines in the ocean deeps without heavy deposition and without mountains, and that there are other geosynclines of great age containing heavy deposits of sediments which have not been folded into mountains. In fact, the precise degree to which there is actual correlation of mountains and geosynclinal sedimentation remains yet to be determined by the students of sedimentation and stratigraphy. When the facts are known it is quite possible that there will be less emphasis on geosynclines of deposition as the principal prerequisite of mountain-building.

Mountain areas of earlier periods have commonly been the

[1] Deeley, R. M., Mountain building: Geol. Mag., new ser., decade 6, vol. 5, 1918, p. 111.

locus of mountain-building in later periods.[1] Some zones of weakness seem to have been permanent through much of geologic history. Many of the principal mountain chains are the result of repeated foldings and upliftings along the same general zones. The younger mountains are in some cases directly superposed on the older ones, but more often they are alongside, thus widening the zone of deformation. In general also the younger ones are on the seaward side of the older ones. Schubert[2] has pointed out for North America that after a geosynclinal prism of sediments is thickened and strengthened by orogenic folding, it does not again become the locus of geosynclinal deposition.

Finally, the localization of mountains is often to be correlated with igneous intrusion, as discussed under the next heading.

Mountain-making and vulcanism. It has already been noted that volcanic extrusions can produce mountains of accumulation, such as volcanic cones. We are principally concerned, however, with tectonic mountains and the intrusions associated therewith. There are many tectonic mountains without evidence of intrusion, but so many are connected with intrusion that there is supposed to be some genetic relation.

Many mountain ranges have igneous cores of a granitic nature. Some of these are merely parts of an old igneous basement, later folded with the superposed sediments, and exposed by erosion. Such igneous rocks clearly are not the cause of the folding — they are merely exposed by the folding.

The igneous cores of other mountains are clearly intrusive, and partly, at least, the cause of the folding, as shown by the zonal distribution of folds, cleavage, and metamorphism about them. The magma may irregularly invade the strata and crumple them, or may follow certain horizons in the manner of laccolithic intrusions, as in the classic Henry Mountain

[1] Willis, Bailey, Some coast migrations, Santa Lucia range, California: Bull. Geol. Am., vol. 11, 1900, pp. 417–432.

[2] Schuchert, Charles, Presidential address, Geol. Soc. Am., Ann Arbor, Mich., 1922.

district described by Gilbert.[1] The sediments are then merely flexed to make room for the magma.

In the pre-Cambrian shield of North America there are immense granitic batholiths of at least three different ages, between which the more or less schistose remnants of overlying rock wind their way, with a more or less mesh-like distribution. These batholiths clearly represent the eroded stumps of mountain ranges. This has always been assumed, but the evidence has recently been well presented by Collins.[2] In part these batholithic intrusions are the folded basement rocks lying unconformably beneath the presently adjacent sediments. In part they are intrusives which invade the sediments and have been largely responsible for their folding. In such cases the zonal arrangement of folds, cleavage and metamorphism around the intrusive batholiths shows beyond reasonable doubt that the intrusion has actually exerted pressure, which is responsible in large part for the adjacent structures. It is entirely possible that there were preëxisting structures which were followed by the intruding magma, but the earlier structures are more or less completely masked by the secondary structures produced by the intrusion itself.

An interesting feature of these batholiths is the frequent inclusion of diversely oriented fragments of schist from the adjacent walls, usually with sharp boundaries, but sometimes with reaction rims, which clearly have been caught up in a molten mass in which the pressures have been essentially hydrostatic. The schist has all the metamorphic and structural characters of the contact zone which has been formed by the intrusion. The writer is inclined to interpret these facts as meaning that the deformation and metamorphism of the surrounding rocks occurred while the magma was still in a molten state, and that the resulting products of meta-

[1] Gilbert, G. K., Report on the geology of the Henry Mountains: U. S. Geog. and Geol. Survey Rocky Mtn. Region, 2d ed., 1880.

[2] Collins, W. H., Physiographic history of northeastern Ontario: Jour. Geol., vol. 30, 1922, pp. 199–210.

morphism were occasionally stoped off and included in the still molten mass.

It does not follow that all batholithic and laccolithic intrusions have caused folding, for cases are known where the preëxisting structures have been little disturbed, as if the magma had made its way, not by crowding the wall rocks, but by stoping them. Extensive intrusive masses in the Bingham district of Utah do not seem to have essentially changed the preëxisting strike or dip of sediments.

In general it may be said that some intrusions probably take advantage of preëxisting structures, or make their way by stoping or assimilation, without structural disturbance of adjacent beds, and to that extent their localization in mountain ranges is a result or incident rather than a cause of mountain-building. But it is also certainly true, particularly for the plutonic intrusions, that they themselves may cause new folding or greatly accentuate the old folding. During the intrusion compressive stresses seem to prevail. With the subsequent cooling of the magma there is contraction, resulting in tension and relaxational movements, — which are likely to manifest themselves mainly in faulting, but to some extent also in folding.

A favored theory of vulcanism is that it is due to release of pressure, which in turn may come through faulting or through the arching and lifting of folds. The weakened structural condition is thought to be the primary condition which thereafter localizes both magmas and orogenic stresses. This is said without implication as to the source of the magmas. Geologists are not agreed as to the extent to which a magma may result from local liquefaction in the zone of orogenic disturbance itself or may come from more distant source below. The common association of both mountain-making and vulcanism with areas of deep, geosynclinal deposition, has suggested that the sediments in the lower part of the geosynclines themselves, together with the basements upon which they are deposited, have softened and fused as a result of loading

and sinking. This is commonly referred to as the hypothesis of *subcrustal fusion*. There are many general considerations which favor the belief that this process is an important one in plutonic intrusion, but it is difficult to find decisive field evidence for it. In some geosynclines of heavy deposition, the bottom layers and the underlying basement are preserved without the slightest trace of fusion. In others, where there are intrusive batholiths, it cannot be said to what extent these are merely the result of geosynclinal down-warping into a hotter zone, and to what extent they are due to the access of new energy and materials from below.

Brouwer [1] has made an interesting attempt to correlate vulcanism with deformation in the Dutch East Indies, and concludes that where the crust is most thickened by overturned folds and underthrusting, nearest the Australian continent, vulcanism has been stopped by the thickening of the crust, thus closing off the conduits.

Chamberlin [2] has called attention to the fact that mountains formed by movement of a thick shell seem to be more commonly associated with vulcanism than mountain-making movements which affect a thin shell, although in the latter case some vulcanism may be present in the deeper portions of the mass. Thick-shelled movements tap a deeper source and are more likely to reach the zone of potential liquefaction of rocks.

A highly significant relationship between mountains and the density of igneous masses has been established by H. S. Washington (see pp. 299–300). The igneous rocks of the higher elevations are lighter than those of depressions. There is close parallelism between the distribution of density caused by the distribution of igneous rocks and that determined by isostatic investigations (see Chapter XIII). It would appear that igneous rocks come to rest at elevations more or less de-

[1] Brouwer, H. A., On the non-existence of active volcanoes between Pantar and Dammer (East Indian archipelago), in connection with the tectonic movements in this region: Proc. Kon. Ak. van Wetenschappen te Amsterdam, vol. 21, 1917, Nos. 6 and 7, pp. 795–802.

[2] Chamberlin, Rollin T., Vulcanism and mountain-making: a supplementary note: Jour. Geol., vol. 29, 1921, pp. 166–172.

termined by their densities, though with modifying structural limitations such as strength of cover, and that mountains with igneous cores owe their origin at least partly to the fact that lighter igneous materials have been concentrated there. In this case the origin and location of the mountains go back to whatever conditions determine the distribution of magmas.

Surface extrusions of lava from volcanoes and fissures are also common, though not general, accompaniments of mountain-making. Vulcanism of this kind is clearly an incident or after-effect of mountain-building and not a primary cause. It may be merely the surface expression of an activity which far below the surface is causing important deformation. Extensive surface vulcanism associated with mountains is often associated with earthquake shocks and recent changes of level, and may be regarded as evidence that mountain-making forces are still active.

The process of mountain-making. Under this heading we may attempt to outline the sequence of mountain-making events to be inferred from the conditions presented on preceding pages. This problem is so much involved with the question of the general causes of earth deformation that some phases of it are left to the discussion of that subject in Chapter XV.

A common, though not universal, condition seems to be heavy geosynclinal deposition along a shore line. This is accompanied and followed by down-warping, folding, and faulting, usually resulting in shortening in a direction normal to the axis of the geosyncline. Both vertical and horizontal movements are ordinarily involved, and often the evidence is that these go on simultaneously. However, there may be uplift without folding or shortening; also there is evidence that folding and shortening are not necessarily contemporaneous with uplift. It is supposed that the Andes and Himalayas were uplifted following the major folding. The detailed history of most mountain ranges shows a long succession of episodes of folding, faulting, and uplift, either singly or in combination.

Vulcanism may accentuate and complicate the deformation, particularly where the deformation is deep, though it is not clear how far it may be cause and how far effect of deformation. Whatever the relation may be, magmas of low density rise higher than magmas of high density, with the result that the cores of many of the higher mountains are dominantly of granitic type and their elevations proportional to their lightness.

As a combined result of tectonic disturbances and vulcanism the mass is greatly thickened and elevated. It remains only for erosion to carve the surface of the elevated block, to complete the mountain as we now know it.

The process is not a slow and uniform one, but its rate varies from time to time; it is periodic. The successive movements are recognized by their relations to unconformities, to peneplains, and to vulcanism. When a mountain movement has started in one place, later movements are likely to be localized either there or near at hand. There is increase in complexity of original structures, and at the same time a widening of the deformed zone. Usually this widening takes place toward the adjacent ocean.

Of course the mountain-making movements of successive periods are not always localized in the same zones. The great movements of the Tertiary period, for instance, mainly affected zones other than those involved at the close of the Paleozoic. Mountain-making involves a thickening and strengthening of the deformed masses, and beyond a certain point stresses may find easier relief in a new area than in one of older deformation.

It is possible that the movements in some ranges have been less intermittent than would be indicated by the time markers available, because of the fact that the erosion represented in unconformities has removed large masses of material, together with their evidences of deformation. Where an older series has been folded and eroded and a newer series has been deposited unconformably upon it and then slightly flexed, we

are apt to say that there was a period of folding preceding the deposition of the new series and another one following. If, however, the area had happened to remain under continuous deposition below sea level, without unconformity, it is entirely possible that the sediments might have shown evidence of progressive and more or less continuous movement throughout the period of erosion on the land area. Brouwer [1] thinks there has been more or less continuous movement from late Mesozoic to the present time in the Dutch East Indies, which is expressed in the areas which have remained under continuous geosynclinal deposition, while on the erosion surface of the islands there is evidence only of an earlier movement and of a distinctly later movement, as represented by the warping of recent coral strand-lines.

Mountain-making movements are both vertical and horizontal. They are by no means uniform in distribution or speed, whether considered in a vertical plane or horizontal plane. One part may move faster than another above or below or at the side, thus developing a drag structure. The bending of geosynclinal axes by horizontal movements along vertical planes normal to the axes have been demonstrated by Brouwer [2] in the Dutch East Indies.

The movements involved in mountain-making are both by rock flowage and rock fracture,— producing folds, faults, joints, and cleavage, with extremely heterogeneous distribution. It is common to assume that the main slipping movement below is by rock flowage, expressed by cleavage and folds, whereas at the surface the movement is mainly by fracture. Folding, however, may be accomplished by fracture as well as by flowage, and hence there is no safe reason to assume that flowage is more prevalent deep below the surface. The vertical distribution of rock structures is discussed in various other places in this book.

[1] Brouwer, H. A., The major tectonic features of the Dutch East Indies: Jour. Wash. Acad. Sci., vol. 12, No. 7, 1922, pp. 172–185.

[2] *Loc. cit.*

Plateaus

Great plateaus like those of western North America and Thibet are ordinarily regarded as mainly the result of vertical uplift, affecting a great thickness of the earth's shell. The boundaries are often marked by vertical faults or by monoclinal flexures in the manner of *horsts* (see p. 67). However, there are also thrust faults and folding, indicating horizontal compression; and in our present state of knowledge it cannot be said how far vertical uplift dominates over horizontal compression in producing these forms. Also, plateaus are by no means single unit blocks of uplift or compression. A plateau may contain several blocks elevated, tilted, and compressed, more or less independently.

Complementary to plateaus there are depressed continental areas with very much the same structure as plateaus.

Sounding of the ocean indicates the presence there of uplifted platforms of a plateau type, together with the broad depressed basins.

Looked at broadly, plateaus and depressed basins may be regarded as partaking of the structures both of mountains and of continents. They are intermediate units of structure, perhaps resulting from the same forces and processes.

Continents and Ocean Basins

General account. Continents and ocean basins may be forms partly inherited from the irregular piling up of heterogeneous material during the growth of the earth, but to some extent they are the results of diastrophism and have acted more or less as units in deformation. A study of the stratigraphic record shows that continents have been uplifted and depressed at various times with relation to the ocean. The distribution and structure of mountains show that there has been crowding between continental and oceanic segments.

Roughly two-thirds of the globe is covered by sea and one-

third by continents. Three-fifths of the continental area is in the Northern Hemisphere. The maximum vertical range between the highest elevation of land and the lowest depth of the ocean is twelve miles. The average difference is three miles. If the continents were graded down and the basins graded up to a common level, according to Chamberlin [1] this level would be about 9,000 feet below the ocean surface. With reference to this level the continents stand up relatively about two miles and the basins are sunk below about one mile.

The continental areas are known to be lighter and therefore of larger specific volume than the sub-oceanic masses, and the two are known to be in a state of balanced isostatic adjustment (see p. 296).

Irregularity, both vertical and areal, is the most apparent feature of the distribution of continents and oceanic basins. If minor irregularities are overlooked, continents may be regarded as of rudely triangular form and the oceanic basins as quadrangular or polygonal in form.

This distribution is taken by Chamberlin [2] to accord with the idea that the earth is a body of somewhat heterogeneous composition which during its growth has shrunk in irregular segments, and that the great negative elements of the earth, represented largely by sea areas — the master segments — should be expected to have polygonal outlines corresponding to the primary place assigned them; that the smaller positive segments or continental areas left between these major segments might be expected to have triangular outlines, or at least fewer angles than the major controlling segments.

A somewhat less obvious pattern is assumed in the so-called tetrahedral theory of the earth. A tetrahedron is a solid body of a geometric form, without reëntrant angles which possesses the greatest possible surface for a given volume. On the hypothesis that the earth's interior is shrinking more rapidly than its shell, it has been inferred that the shell would

[1] Chamberlin, T. C., and Salisbury, R. D., Geology: Henry Holt & Co., New York, vol. 1, 1904, p. 523.

[2] *Loc. cit.*, pp. 521–522.

tend to maintain the largest possible area of surface and therefore might take on tetrahedral lineaments. Continental areas and mountain chains would then correspond roughly to the angles and corners of the tetrahedron. If a tetrahedron is stood on one of its corners and this point called the south pole, the three upper corners and angles are supposed to correspond to the land areas surrounding the north pole. The three angles extending down toward the south polar point correspond to the continental ridges of South America, Africa, and Australasia. The dominance of land area in the northern half of the earth would accord with the dominance of projections in the upper half of the tetrahedron. It is needless to say that this comparison requires some imagination. It is cited merely as illustrative of the several hypotheses offered. Equally good comparisons have been made with other geometric forms.

The influence of preconceived hypotheses of the origin of the earth is clearly perceptible in generalizations as to the actually existing distribution of continents and sea basins. Different generalized patterns may be superposed on earth lineaments, and each will have many points of accordance. Which of the patterns, if any, has real significance one cannot yet positively say.

Nature of movements affecting continents and sea basins. All of the movements, orogenic and epeirogenic, producing structures described in previous chapters, of course affect continents and sea basins. Their cumulative effect probably accounts for a considerable part of the movements on a continental scale. Our interest in net results leads us to discuss these larger movements as if the continents and ocean basins had acted more or less as units. This large view of the situation brings out certain features of earth deformation which are not apparent in the detailed structures.

The sedimentary record shows that the present continental areas have risen and fallen many times with reference to the level of the ocean. They have alternately suffered submer-

gence and emergence. The fact of emergence of a continent is not necessarily evidence of actual uplift with reference to the center of the earth. Different parts of the earth's surface might all be moving toward the center, but if the rates be different, the part that is lagging behind might seem to be uplifted. The movements are continuing to the present time, as shown by historic changes in sea levels and old beaches.

Fragments of the sedimentary record which are available from under the sea likewise indicate vertical movements. Areas are submerged which were once clearly land. Shallow water deposits now lie in deep water, and deep water deposits have been brought to shallow depths.

Other evidence of uplift and depression which is less direct, but conclusive, is derived from conditions of isostatic adjustment, described in Chapter XV.

The nature of the sediments which have been deposited over continental areas during periods of submergence indicates that the submergence was not very deep. Most of the sediments were deposited at depths less than a thousand feet; the nature of the contained organic remains indicates that they were deposited within the depth to which light is known to penetrate. Abundant deposits of a deep sea type are notably absent. In short, continents have remained continents, notwithstanding their temporary shallow submergence. The ocean deeps seem also to have remained more or less permanent during geologic time, so far as can be told from the scant evidence available. Intermediate areas, like continental shelves, and minor submerged platforms within the ocean, have obviously been both above and below the ocean at various times. So far as known, a fairly correct generalization is that the present continental areas and the present oceanic depths have in the main been such through geologic time. Geologists are not all agreed on this conclusion, but the preponderance of opinion seems to be as stated.

So obvious is the evidence of vertical continental movements that perhaps too little attention has been paid to hori-

zontal movements indicated by a different kind of evidence. The localization of mountains along the borders of continents and along old geosynclinal shore areas seems to indicate that the borders of great continental and oceanic segments are areas of lateral crowding. The shortening involved in great mountain chains implies horizontal drift of considerable surfaces of the globe. Horizontal movements in the Dutch East Indies are thought to be at least ten times the vertical movement in extent.[1] The great zone of Tertiary mountain-building along the southern part of the Eurasian continent was long ago interpreted as marking an important tangential movement of the crust southward from the pole.

During historic times there have been horizontal migrations of land areas, as shown by geodetic surveys. Thus, north Greenland is said to have moved westerly at the rate of 9 to 32 meters per year since 1823. Lateral movements of geodetic stations have been noted in connection with the study of California earthquakes. European and North American observations are reported to show a slight decrease of latitude with lapse of time. Observations are not sufficient to allow sound generalizations as to directions of present movement for the earth, though some investigators seem to discern a drift westerly and toward the equator.

Facts of this kind and many others of similar import have been used as the basis for highly speculative hypotheses of the origin of continents. Taylor [2] ascribed the origin of the earth's plan, including the distribution of continents and mountain chains, to a drift away from the poles, arising from increase of oblateness. Considerably later Wegener [3] postulated that the continents are light, salic masses floating on a molten ferric or basaltic substratum, and that under the influence of

[1] Brouwer, H. A., The horizontal movement of geanticlines and the fractures near their surface: Jour. Geol., vol. 29, 1921, pp. 560–577, and personal communication.

[2] Taylor, F. B., Bearing of the Tertiary mountain belt on the origin of the earth's plan: Bull. Geol. Soc. Am., vol. 21, 1910, pp. 179–226.

[3] Wegener, A., Die Entstehung der Kontinente und Oceane: Die Wissenschaft, 1920.

the earth's rotation the continents of the Northern Hemisphere are moving westerly and toward the equator; that the continents have been separated by this drift; that, for instance, South America has broken away from Africa, as indicated by their complementary outlines, the close relations of faunas prior to the supposed break, etc.

Movements of this kind imply slipping over a surface with exceedingly low gradient, and therefore with a high degree of mobility, and there are many mechanical difficulties involved.

There is still another kind of lateral movement from the continents toward the ocean — by ordinary erosion and transportation of waste, cutting down the higher parts of the continents and spreading them along the margins, thereby encroaching on the deep-sea areas.

CHAPTER XI

EARTHQUAKES

Nomenclature

An earthquake is a shaking of the earth. *Seismology* is the science of earthquakes. Small vibrations continuing for hours are known as *microseisms*. These minor, unfelt disturbances, recorded only by delicate instruments, have also been called *cryptoseisms*. Major vibrations are called *macroseisms*. Earthquakes of deep-seated origin are known as *bathyseisms*. The *focus* of an earthquake is its starting point. The *epicenter* or *epicentrum* or *epifocus* is the spot on the surface just over the locus of origin of an earthquake.

Earthquakes as an Effect of Rock Failure

Earthquakes are not in themselves rock structures but may be both cause and effect of rock deformation, particularly fracturing.

Many earthquakes are caused by movement along pre-existing fault planes or agree in position with known faults. Intensity of disturbance is greatest near the fault and decreases away from the fault. Furthermore, nearly all earthquakes are of very shallow origin, suggesting their correlation with an assumed zone of fracture. These facts have led to the generalization that all earthquakes are due to fracturing near the earth's surface. There are other facts, however, which suggest that this generalization should be qualified. There have been great earthquakes, like the Charleston earthquake, affecting large areas, in which there has been no

observed evidence of tectonic disturbance. Oldham[1] calls attention to the fact that in the Indian earthquake of 1897

"there was no single leading fault and zone of maximum intensity of shock, but a complicated network of lines of extreme destructiveness, ramifying over an area not much different from that of England, and extending right across a series of great tectonic features, across the great monocline of the southern face of the Assam range, across that range itself, across the alluvial plain of the Brahmaputra valley, the great boundary-faults of the Himalayas, and probably even across the main axis of elevation of the range."

It cannot be proved that these earthquakes were not due to rupture of some zone beneath the surface, but the relation to rupture is not a simple and obvious one, and Oldham suggests that the shock may be due to some deeper-lying cause, such as sudden changes of volume of the rock mass which may or may not be associated with faults. Even where earthquakes are related to known faults, it may be that both the faults and earthquakes are the results of some underlying cause, and that the faults merely serve to localize the disturbance. If the faulting were not there, there might still be the earthquake, but with different distribution.

It is not certain, also, that earthquakes may not have some relation to sudden deformation by rock flowage. When it is remembered how intimate is the association of fracturing and flowage, both in time and place, and that both are the expressions of the same stresses acting on different rocks, it becomes difficult entirely to exclude rock flowage from consideration in connection with earthquakes. The principal argument against it is that rock flowage is conceived to be essentially a slow, cumulative process — too slow in itself to cause any disturbance.

So-called volcanic earthquakes (see p. 268) may also in part be independent of fracturing.

[1] Oldham, Richard Dixon, The anniversary address of the President: Quar. Jour. Geological Soc. of London, vol. 78, pt. 1, No. 309, 1922, p. lxvi.

The faults actually observed to be related to earthquakes are mainly along vertical or highly inclined planes. The movement along these planes has been vertical, horizontal, and inclined. Low-angle thrust faults doubtless also originate earthquakes but these have seldom been observed in process of formation near the surface, and still more rarely, if ever, definitely connected with earthquakes. On land areas the fault displacements causing earthquakes are usually to be measured by a few feet or a few tens of feet, though it is likely that larger fault displacements of the past also caused earthquakes. Observations taken in sounding and on the breaks in cables following earthquakes seem to show that large segments of the bottom of the ocean have dropped hundreds of feet as a cause of the shocks. In general it may be said that no relation has been worked out between the magnitude of fault displacements and that of the resulting earthquakes.

A highly interesting outcome of the study of the causal relation of faults to earthquakes has been the development of the *elastic-rebound* theory, which is formulated by Reid[1] as follows:

"1. The fracture of the rock, which causes a tectonic earthquake, is the result of elastic strains, greater than the strength of the rock can withstand, produced by the relative displacements of neighboring portions of the earth's crust.

2. These relative displacements are not produced suddenly at the time of the fracture, but attain their maximum amounts gradually during a more or less long period of time.

3. The only mass movements that occur at the time of the earthquake are the sudden elastic rebounds of the sides of the fracture towards positions of no elastic strain; and these movements extend to distances of only a few miles from the fracture.

4. The earthquake vibrations originate in the surface of fracture; the surface from which they start has at first a very small area, which may quickly become very large, but at a rate not

[1] Reid, Harry Fielding, The elastic-rebound theory of earthquakes: Bull. of Dept. of Geol., Univ. of California, vol. 6, No. 19, 1911, p. 436.

greater than the velocity of compressional elastic waves in the rock.

5. The energy liberated at the time of an earthquake was, immediately before the rupture, in the form of energy of elastic strain of the rock."

The displacement at first is within the elastic limit, and this is suddenly translated into a less extensive displacement expressed by permanent rupture. We ordinarily think of the displacement as developed solely by the faulting. The initial strain is supposed to be due to slow, subcrustal earth currents. These might be opposing currents on two sides of the fault plane, or they might be all in the same direction but at different rates of speed when considered either vertically or horizontally. The drag from these currents is supposed to be applied on the underside of the fracturing crust, although this is not a necessary consequence of the hypothesis.

Reid applied the elastic rebound theory to the great California earthquake of 1906, which followed the San Andreas fault in the coast ranges of California, and Lawson [1] has more recently developed the idea still further. Lawson concludes that the cause of the earthquakes along the San Andreas fault is probably a northerly subcrustal flow, expressing itself (1) as a longitudinal strain opposed in direction to the general stress, which is relieved by lowly inclined, deep faults, having strike normal to the direction of stress; and (2) as a transverse strain, due to the unequal distribution of the stress in the horizontal sense, which is relieved by vertical faults having a strike oblique to the direction of stress. The elastic strains which caused the last California earthquake were not developed suddenly or immediately before the shock, but existed to some extent twenty-five and fifty years earlier. These conclusions are based on geodetic surveys made before the earthquake of 1868, those made between 1868 and 1906, the time of the last great earthquake, and those made since.

[1] Lawson, Andrew C., The mobility of the Coast Ranges of California — an exploitation of the elastic rebound theory: Bull. of Dept. of Geol., Univ. of California, vol. 12, No. 7, 1921, pp. 431–473.

The data are not yet entirely sufficient but the continuation of surveys in the future may afford a definite check of this hypothesis.

At present this idea of the development of earthquakes may be regarded merely as a good working hypothesis. The geodetic surveys have been too infrequent to make sure that the accumulation of strain was really slow and uniform. There may have been long periods of quiescence between the surveys and more or less sudden accumulations of strain. Also it is difficult to be sure that the base points, with which the moving points are compared, have themselves really remained stationary.

The idea that sub-crustal creep is the cause of surface fracture and of the localization of earthquakes, is held also by Brouwer [1] for the recently moving geosynclinal areas of present deposition in the Dutch East Indies.

Earthquakes as a Cause of Rock Fracture

Where rocks are already under strain, earthquakes may bring the stresses beyond the breaking point. This has been accomplished experimentally,[2] and presumably many fractures and systems of fractures have been so caused in the earth; but fractures of this origin have not been positively identified. It is highly probable that some of the complementary fractures associated with great fractures which originate earthquakes are secondary results, due to the transmission of the shock from the primary source into a zone already under strain. The after shocks following a great break may well be due to secondary fractures caused in this way. The data relating earthquakes to specific rocks and rock structures are yet only fragmentary.

[1] Brouwer, H. A., The major tectonic features of the Dutch East Indies: Jour. Wash. Acad. Sci., vol. 12, No. 7, 1922, pp. 172–185.

[2] Crosby, W. O., The origin of parallel and intersecting joints: Am. Geologist, vol. 12, 1893, pp. 368–375.

Other Effects of Earthquakes

While rock fracture is ordinarily thought of as the common result of earthquakes, we can hardly eliminate the possibility that they may also under proper conditions produce or accelerate rock flowage and folding.

The maximum shaking and destructive effects of earthquakes appear in loosely consolidated rocks, gravels, and soils which are saturated with water. The reason for this is not entirely clear. It has been suggested that the water affords opportunity for the materials to move easily, and by filling all the pore spaces that it aids in the transmission of the shock.

Further effects are landslides, snowslides, acceleration of glacial motion, water waves on the ocean, various disturbances in surface forms and drainage, changes of level, damage to human structures, vulcanism, and magnetic disturbances. Some of these results may be nearly instantaneous, but others may lag long behind the primary shock. The acceleration of movements of glaciers in Alaska has been observed to follow months behind.[1] It seems inherently probable that slow earth movements may likewise be initiated, though direct evidence of this is still lacking.

Earthquakes and Vulcanism

Volcanic belts lie in or near great earthquake zones. Vulcanism has been accompanied in many places by earthquakes, and vice versa. However, there are many cases of earthquakes not associated with vulcanism, and of volcanic outbreaks not accompanied by earthquakes. The outbreaks at Krakatoa and Mt. Pelee caused only minor vibrations; the California earthquake was not accompanied by vulcanism. There is, therefore, no necessary connection; but since vulcanism is now generally regarded as involving mechanical dis-

[1] Tarr, R. S., and Martin, Lawrence, The earthquakes at Yakutat Bay, Alaska, in September, 1899: Prof. Paper 69, U. S. Geol. Survey, 1912.

turbances of the crust, lessening the pressure upon the hot rock and thereby allowing it to liquefy, it may be reasoned that earthquakes, by disturbing the equilibrium of pressures, may be a local cause of vulcanism. Or both may result from larger earth movements.

A volcanic earthquake is one due to the direct action of the volcanic force or one whose origin lies under or in the immediate vicinity of a volcano, whether active, dormant, or extinct. From a study of earthquakes of this kind Davison [1] concludes:

> "(*a*) That the foci of volcanic earthquakes are situated at a very slight depth below the surface; (*b*) that the foci are usually small and seldom more than 4 or 5 miles in length; (*c*) that the after-shocks, when they occur, originate chiefly within the focus of the principal earthquake; and (*d*) from the discussion of the Etnean and Alban earthquakes, that, while the majority of volcanic earthquakes originate along radial fractures of the mountain, some, and by no means the least important, originate along perimetric fractures."

Davison concludes further that volcanic earthquakes are of tectonic origin insofar as they are due to the growth of faults, but of volcanic origin in that the slips are precipitated by present or past volcanic activities.

Correlation of Earthquakes with Other Natural Phenomena

As listed by Cotton,[2] attempts have been made to relate frequency and periodicity of earthquakes to:

1. Auroral displays and luminous phenomena.
2. Magnetic and electrical disturbances.
3. Planetary influences.

[1] Davison, Charles, Volcanic earthquakes: Jour. Geol., vol. 29, 1921, pp. 97–124.

[2] For a review of literature on these questions and particularly for evidence of the relations between earth tides and earthquakes, see: Cotton, Leo A., Earthquake frequency, with special reference to tidal stresses in the lithosphere: Bull. Seismological Soc. Am., vol. 12, 1922, pp. 47–198.

4. Sun spots.
5. Temperature.
6. Barometric pressure.
7. Wind velocity.
8. Rainfall and snowfall.
9. Seasonal periodicity.
10. Diurnal periodicity.

(4–10: Solar effects.)

11. Ocean tides.
12. Earth tides.

(11–12: Solar and lunar effects.)

13. Other periodicities including latitude variations.

The conclusions reached in these investigations are more or less conflicting, but there is a tendency for seismologists to believe that there is some relation between the time of earthquakes and barometric changes, ocean tides along certain coast lines, and tidal stresses of the earth's solid crust.

Earthquake Waves

We cannot do better than to quote a summary by Reid: [1]

"Nearly all earthquakes are due to the sudden fracture of the rock of the earth's crust, which has been strained by slow earth movements beyond its strength. Strong vibrations are set up at the fractured surface at the time of the fracture, just as vibrations are set up whenever any solid is broken. The earth consists of solid material, which is necessarily elastic, and therefore these vibrations are transmitted through it as elastic waves. There are two kinds of elastic waves: normal waves, where the movement is in the direction of propagation, and transverse waves, where the movement is at right angles to the direction of propagation. These two kinds of vibration advance with different velocities. Their velocities near the earth's surface are about 7 and 4 kms. per second, respectively; but the deeper they penetrate below the surface the faster they go. Therefore, from the characteristics of elastic waves their paths are curved and concave upwards. These important results have been obtained by a study of the time necessary for the two types of waves to pass from the place of their origin to stations at dif-

[1] Reid, H. F., The problems of seismology: Proc. Nat. Acad. Sci., vol. 6, No. 10, 1920, p. 556.

ferent distances where delicate instruments are installed, which record the time of their arrival. It may seem remarkable that the paths followed by the rays and the velocity in different parts of the paths could be determined merely from the time required to arrive at a number of places on the earth's surface: but, by the help of some rather abstruse mathematics, it can be done. One interesting conclusion which can be drawn from the passage of the transverse waves through the body of the earth is that the earth is a solid and not a liquid sphere; for transverse waves can be transmitted only by solid substances."

"There are other waves which are transmitted along the surface of the earth; they must be started in some way when the body waves, mentioned above, arrive at the surface, but we do not know just how near the origin their starting place is."

"In many instances submarine earthquakes give rise to great water waves, which have been known to travel from one side of the Pacific Ocean to the other. It has been the general belief that the first indication of these waves along a coast is marked by the withdrawal of the water; and the great elevated wave follows. Although this order is certainly frequent, it is not general, and in many cases the elevated wave is first to appear."

What is above called a normal wave has also been called a *compressional* or *primary* or *longitudinal* wave. The transverse wave is sometimes referred to as a *distortional* or *secondary* or *shear* wave. The circumferential waves are undifferentiated. These waves follow in the order named and constitute a three-phase record. A common statement in older seismological treatises is that this record is more or less confused within a distance of 700 miles of the earthquake origin, or 10° from the epicenter; but there are recent examples of clean-cut separation of the three phases at much less distances, in one case 29 miles.[1] Beyond 110° from the epicentrum they become confused and dampened. It is not known whether the waves passing through the deep core of the earth are the compressional or the transverse waves, or both.

[1] Macelwane, James B., Some seismological evidence that is not evident: Science, vol. 56, 1922, pp. 478–480.

Seismographs are instruments for the detection and measuring of earthquake waves. They are made in a variety of forms but are all essentially devices for determining more or less independently the three principal components of the waves, that is, the vibrations in three mutually perpendicular planes.

Method of Locating Origin of Earthquakes

The origin of earthquakes has been found to be shallow wherever it has been possible to determine the directions of emergence of earthquake waves, either from instrumental observations or from the study of the destructive results of earthquake shocks. The rate of variation of intensity of the disturbance with reference to its epicenter gives some notion of the depth of origin. The nearer to the surface the more nearly do the variations shown at the surface correspond to the variation in intensity from the focus. Nowhere have these determinations indicated a depth of origin greater than twelve miles, and usually less. The very fact that an earthquake shock is usually so well localized at some spot on the earth's surface, that there is some one zone which may be regarded as the locus of activity, is evidence that its origin is not far below the surface. Usually the locus is not a point but a line or elongated zone, corresponding to the position of a fault plane. This may extend horizontally tens of miles or a few hundreds of miles.

In the area most affected by the quake, the origin is located by the intensity of the shock and by noting the direction of emergence of the waves. The area most affected is usually roughly oval or elliptical, and within it there is usually a line or spot at which the intensity of the shock is clearly at a maximum. It is assumed that near the origin the waves are both transverse and compressive, that both shearing and tensional stresses are set up in the structures affected, and that the breaking strength is first surpassed by tensional stresses, the dominant one of which would be normal to the

direction of transmission of the wave. Hence fracture planes in buildings are regarded as due to tension, and, therefore, normal to the path of the wave. The plane of fracture is best determined at the corner of a building. Lines drawn normal to these fracture planes in widely distributed areas may tend to converge in a point, or plane, which is then regarded as the origin of the quake. This method is of doubtful value, because the attitude of fractures is much influenced by local conditions, and it is difficult to prove that they are tensional.

The location of the earthquake from more distant points is accomplished mainly by noting the difference in time of receipt of the first arriving normal or compressional waves and the later arriving transverse waves. The greater the difference in time between the receipt of the two, the greater is the distance from the point of origin. The rates of each of these kinds of waves being known, the difference in time of their arrival affords basis for calculating the distance. At any one point of observation, distance, not direction, is determined. It needs observation of distance from three points to determine by the intersection method the locus of origin. As a matter of fact a good guess is often made from one or two stations. If at one station the earthquake is estimated to be 2,000 miles distant, the drawing of a circle with this radius around the station may intersect only a single zone where earthquakes are known to be frequent, and this is regarded as the possible locus. If there are two stations, the locus is known to be at an intersection of the circles drawn from these two points, and the choice of the point may be determined by its relation to zones of common earthquake disturbance.

Earthquake Zones

The distribution of recorded earthquakes corresponds with zones of more or less intense deformation or vulcanism, or both. In a very general way there are two great earthquake zones: the so-called Mediterranean zone or belt, passing

through the Himalayas and eastern China, from which have started 53 per cent of the recorded earthquakes; and the Pacific belt, bordering the Pacific basin, in which have originated 41 per cent of the recorded earthquakes.[1] More specifically, earthquakes are likely to follow along the margins of

FIG. 99. Maps showing, in black, the principal earthquake regions of the world. After Montessus de Ballore.

continents or of smaller areas of great relief, along mountain chains, especially of recent origin, along volcanic belts, along margins of two areas differing considerably in density, as for instance in the zone of the Messina earthquake, and along areas where there are great irregularities in distribution of the

[1] Montessus de Ballore, F. de, Les Tremblements de Terre, Paris, 1906.

earth's magnetism. It has been ascertained that earthquakes have been especially numerous in the geosynclines of Mesozoic rocks. As many of these rocks have been folded into mountain ranges in comparatively late geological time, this is only

FIG. 99 (cont.).

a specific case of the abundance of earthquakes in mountains of recent origin.

The distribution of earthquakes under the sea is still imperfectly known.

Condition of the Interior of the Earth as Inferred from Earthquakes

The passage of transverse waves through the earth to great depths (at least 110° from the origin) has proved that the earth is solid to this depth, for transverse waves cannot exist in a liquid. The velocity of transmission depends on the ratio of the elasticity to the density of the medium. The greater the elasticity, the greater the velocity; the greater the density the less the velocity. Since the velocity of earthquake waves is increased with depth below the surface, elasticity of the earth may be inferred to increase faster than the density. Elasticity is not the same as rigidity, but high elasticity is usually taken to imply high rigidity.

Earthquake waves passing through the core of the earth beyond 110° from their origin become confused and dampened, and seismologists are not agreed on the extent to which the core is crossed by longitudinal or transverse waves or both. On the assumption that transverse waves do not cross the core it has been argued that it may be liquid. By others the evidence is taken to favor the idea that the core is made of iron. Still others regard the evidence as satisfactorily explained by the assumption that the core of the earth is not different in composition, but is different in density because of compression. Experiments on the compressibility of rocks which have been conducted under pressures corresponding to a depth of 50 kilometers indicate a certain amount of compressibility; but the extrapolation of these results to pressures corresponding to greater depths fails to indicate sufficient increase of density to account for the known mean density of the earth, thus favoring the assumption that the core is made of some different material like iron.[1]

Most seismologists and geophysicists suppose the core to be rigid. The rigidity of the earth as a whole, in its behavior

[1] Williamson, Erskine D., Change of the physical properties of materials with pressures: Jour. Franklin Inst., vol. 193, 1922, pp. 491–513.

under rather sudden stresses, is known from other than seismological evidence (see pp. 292–293), to be higher than that of steel.

Reid[1] calls attention to the fact that fully 95 per cent of the energy of an earthquake shock comes to the surface within the hemisphere having the origin as its pole, and concludes that the data on its transmission for greater distances, through the core of the earth, are too imperfect to yield reliable deductions.

There have been various attempts to deduce the distribution of density and rigidity within the earth from the form of the transmission curve of earthquake waves, and several hypotheses have been proposed to the effect that the earth is made up of a core of high density, surrounded by one or more sharply differentiated shells of lower density in which the velocity suddenly changes. The hypotheses do not agree on the thickness or number or density of the shells.

The doubtful fact (see p. 270) that the separation of normal, transverse, and circumferential waves becomes distinct only at distances at 10° of arc, where waves have penetrated at least 100 kilometers below the surface, has been interpreted to mean the existence of an isotropic medium at a comparatively shallow depth beneath the surface, in contrast to the known heterogeneous crystalline character of the earth's surface, and, therefore, to favor the assumption that below this depth the earth is in different physical condition than at the surface. It is quite possible, however, that material may be homogeneous in a large way and yet possess crystalline heterogeneity like the surface rocks.

Prediction of Earthquakes

No means have yet been found of predicting the time and place of earthquakes with any considerable degree of success.

As to place, there is the probability that earthquakes will be confined to certain broad zones in which they have com-

[1] Reid, H. F., The problems of seismology: Proc. Nat. Acad. Sci., vol. 6, No. 10, 1920, pp. 557–559.

monly originated in the past. Within an earthquake zone the records seem to show that a great disturbance at one locality may mean that the next disturbance is to be looked for in some other part of the belt. There have been too many exceptions to this, however, to establish the rule. When one notes the widespread distribution of faults over the earth's surface, many of them doubtless accompanied in their genesis by earthquakes, and considers the possibilities for faulting in the geologic future, predictions as to the localization of earthquakes, based on the meager records of historical time, cannot be accepted with any great confidence. It is to be remembered also that great earthquakes sometimes occur where there is no surface evidence of faulting, past or present.

Reid [1] suggests that it is not impossible that the location of preliminary shocks might serve to block out the fault along which a great rupture was about to take place. Little is known about the preliminary ruptures, except that they probably occur.

Much effort has been given to the attempts to establish the periodicity of earthquakes for different regions, and to correlate the periods with the tide-producing stresses, with variations of latitude, with climatic periods, and with slow accumulations of strain by earth currents. The elastic rebound theory is now much thought of as a possible means of predicting earthquakes. If there is a uniform and slow accumulation of strain which can be measured, and if it can be established for any locality that there is some periodicity in relief from this strain by fracture, this becomes a possible basis of prediction.

As yet these attempts have not been successful, perhaps because of the insufficiency of data. Gilbert [2] calls attention to the fact that many attempts at working out the periodicity of earthquakes are apparently successful because the great frequency of earthquakes of some magnitude furnishes examples for any time-system postulated.

[1] Reid, Harry Fielding, The problems of seismology: Proc. Nat. Acad. Sci., vol. 6, No. 10, 1920, p. 559.

[2] Gilbert, G. K., Earthquake forecasts: Science, vol. 29, 1909, pp. 121–138.

CHAPTER XII

UNCONFORMITY

Significance of Term

Contiguous formations are said to be unconformable where there is evidence of an erosion interval of some magnitude between their periods of formation or evidence of cessation of

Fig. 100. Horizontally bedded limestone, resting unconformably on vertical beds of Proterozoic quartzite. Box Canyon, near Ouray, Colo. After R. T. Chamberlin.

deposition between them. In either case there is loss of part of the geological record. The term *unconformity* is sometimes used to indicate primarily the physical discordance; sometimes it is applied principally to the time interval implied by

the discordance; it usually implies both. Other terms used for unconformity are *disconformity* and *nonconformity*. The term nonconformity is sometimes used where the upper and lower formations are visibly different, or have different attitude, and disconformity where the two formations are in parallel position.[1]

The deformation of rocks, with which this book is mainly concerned, is only one of the factors to be considered in unconformity. Stratigraphy, physiography, and paleontology are others, — in fact, adequate understanding of the significance of unconformity involves the widest range of geological knowledge. The subject is treated here principally in its relation to structural geology, and not in the broader sense that is required for an understanding of its full significance.

Identification of Unconformity

Physical evidences of unconformity are:

(1) *Evidence of erosion*, between older and younger formations, which may be parallel or non-parallel.

(2) *Differences in metamorphism.* Stratigraphically lower rocks may have suffered so much more metamorphism than

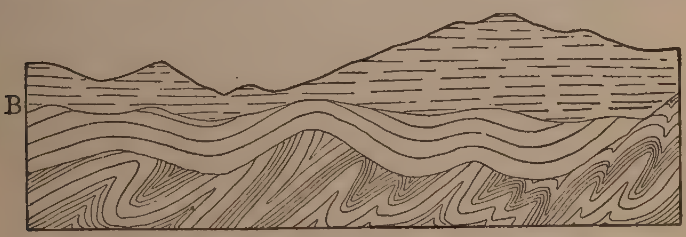

FIG. 101. Ideal sketch to illustrate unconformities. After Spurr. *A*. Earlier line of unconformity; *B*. Later line.

overlying beds of similar lithology as to indicate the probability of a time interval between them. Original differences in

[1] Pirsson, L. V., and Schuchert, Charles, Textbook of Geology: John Wiley and Sons, 2nd ed., part I, 1920, p. 311.

lithology also influence the nature and extent of metamorphism, causing sharp contrasts in metamorphic effects in contiguous conformable beds.

(3) *Difference in deformation.* Stratigraphically underlying rocks may be folded or cracked or may be schistose as a result of flowage, while these features may be less conspicuous or lacking in upper beds of similar kinds, indicating a time interval between their periods of formation. This criterion must be carefully used, for the differences in deformation may be due simply to varying competence of the different beds in a conformable sequence.

(4) *Difference in number of igneous intrusions.* Stratigraphically underlying beds may be intruded by igneous rocks which have not intruded the upper beds. This may not in itself be evidence of unconformity, but may confirm other evidences of the existence of an erosion interval between lower and upper beds. If the igneous rock in the lower beds is plutonic rock and appears on the contact erosion surface, it is evidence that the erosion interval has been of sufficient duration to allow of the removal of a great thickness of rock.

(5) *Basal conglomerate,* in the upper beds, carrying fragments from the rocks beneath the contact surface. If this conglomerate contains a variety of fragments derived from a considerable area, it is more significant of a time interval, perhaps, than a conglomerate made up of fragments entirely like the immediately underlying rock. However, if the underlying rocks are homogeneous over great areas, the overlying basal conglomerates may show a marked homogeneity of fragments. Intraformational conglomerates are sometimes formed by exceptional storms or other causes, in the course of a continuous deposition of sediments. Such conglomerates mark no erosion interval of magnitude and do not signify unconformity (see p. 111).

While a basal conglomerate indicates unconformity, the absence of such a conglomerate does not prove conformity, for students of sedimentation now find many conditions under

which sediments may be deposited unconformably on older surfaces without intervening conglomerates. The base of the Paleozoic in the Mississippi Valley as a whole is remarkably free from basal conglomerates, except near monadnocks on the old pre-Cambrian peneplain. Some of the great unconformities of the pre-Cambrian exhibit similar conditions. In fact, if the writer were to discuss basal conglomerates solely from the standpoint of his own field observation, he would state that they are the exception and not the rule; that over great, flat surfaces of deposition conglomerates are either entirely absent or very small, thin, and inconspicuous; and that the large, conspicuous basal conglomerates usually mantle around some sharp elevation projecting above the general plane of deposition. The commonly assumed sequence above an unconformity is conglomerate, grading into coarse sand, and this in turn into fine sediments. It is not at all uncommon, however, to find at the base of the overlying series a few feet or a few tens of feet of finely bedded argillaceous sand rocks, which may rest directly upon the underlying basement or be separated from it by a thin conglomerate. When the beds are later disturbed and exposed to erosion, this relatively soft material weathers rapidly, with the result that the unconformable contact often lies in a depression without exposures. Where the underlying formation is of hard rock, such as quartzite or granite, the best chance for finding a basal conglomerate is along the eroded face of the older series near the contact, where patches of the overlying conglomerate adhere to the harder, older rocks.

(6) *Field relations* and areal distribution of rocks may indicate an unconformity even where actual contacts or other evidences are lacking. For instance, a continuous bed of quartzite lying alongside of a heterogeneous group of rocks with irregular distribution would in itself suggest unconformity between these rocks and the quartzite. It is frequently possible from a preliminary study of maps showing areal distribution of lithologic types to infer possible unconformities, and

to direct further field work with much greater effectiveness than would otherwise be possible.

(7) *Difference in lithology;* as, for instance, where a sedimentary rock rests upon an igneous rock without intrusive relations.

(8) *Hiatus in the fossil record* between successive beds.

(9) *Absence of certain beds* from a known sequence may indicate local unconformity, though not always so.

Commonly the greater number of these criteria can be combined in working out unconformity. One line of evidence can usually be substantiated by others. No single criterion is likely to be conclusive.

Interpretation of Unconformity

Unconformity represents a lost interval not otherwise recorded at that place. This lost interval may involve (*a*) a cessation of deposition, usually involving emergence, and often accompanied by deformation of the rocks; (*b*) denudation, usually by subaerial processes; (*c*) resumption of deposition, usually following submergence, but often by terrestrial processes.[1]

The description and interpretation of unconformity is usually based mainly on the assumption that the overlying sediments are subaqueous, and that the preceding interval of erosion must necessarily imply emergence or uplift from the sea. It is recognized, however, that great sedimentary formations may be deposited on land areas with unconformable relations to underlying rocks, in which case the period of erosion does not necessarily involve uplift. Furthermore, unconformity may be caused by mere cessation of deposition without uplift. Just as erosion stops at the base level stage, so there is cessation of deposition at times and places under water. Resumption of deposition at these places creates an unconformity marked by a lost interval of deposition, if not by structural discordance.

[1] Blackwelder, Eliot, The valuation of unconformities: Jour. Geol., vol. 17, 1909, p. 290.

The appraisal of the value of an unconformity requires much care. The terms "great" and "slight," frequently applied to unconformity, express the value very crudely. By "great unconformity" may be meant one in which there is a prominent discordance of structure, or one indicating the absence of great thicknesses of strata, or a long lapse of time, or any combination of these features. Usually it is intended to imply that the discordance is pronounced and that there is a great loss of record. It is desirable, wherever possible, that these factors be discriminated, even though their quantitative value cannot be determined.

The study of unconformities broadly as continental features is of significance to structural geology as indicating the major warpings and oscillations of the continents with reference to the sea. The continents from the beginning of the geological record have always in some part stood above water, have in some part been undergoing erosion, and therefore lack a complete record of deposition. By migrating from place to place during continued movements, animals might conceivably have lived continuously on the erosion surfaces which marked unconformities in the geologic record. Thus it appears that in one sense unconformities are continuous physically and chronologically, but that they shift back and forth across the continents with successive oscillations and inundations. Any localized unconformity is represented somewhere else by a continuous record of deposition. As Blackwelder [1] states it:

> "The entire geologic record, then, is not to be conceived of as a pile of strata, but as a dovetailed column of wedges, the unconformities and rock systems being combined in varying proportions. The former predominate in some places and periods, while the latter prevail in others."

In the study of unconformity we are so much concerned with the break in the sedimentary record that we may fail to realize that unconformity also signifies a break in the

[1] *Op. cit.*, p. 299.

deformational record. Erosion takes away sediments in which earth movements are registered. Structures representing movements before and after the unconformity are assigned to separate and distinct periods, with the concealed assumption that there was no movement in the time between, represented by the lost record. A study of movements in the sediments

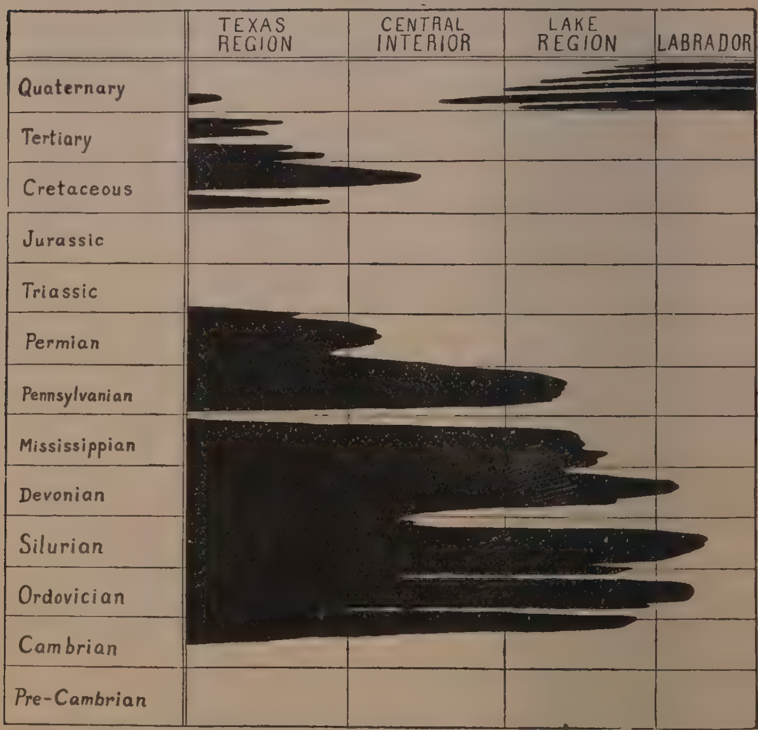

FIG. 102. Diagram of an unconformity with lateral extensions and restrictions. After Blackwelder.

The extent and duration of the principal periods and areas of sedimentation, with their corresponding rock systems, are shown in solid black. The white, on the other hand, denotes the time and extent of erosional conditions and corresponding unconformities.

of the same age deposited continuously in an adjacent geosyncline may indicate that movements of a mountain-building nature were continuous and not recurrent (see pp. 254–255).

CHAPTER XIII

THE STRENGTH, DENSITY, AND VOLUME CHANGES OF ROCKS[1]

The foregoing chapters have dealt with evidences of rock failure, with only incidental reference to their strength, density, and volume changes. It is desirable to bring together some of the known facts regarding them, as a background for the general consideration of the causes and nature of diastrophism to be taken up in the following chapters.

The Strength of Rocks

When we speak of the strength of rocks we usually have in mind the capacity of rocks to resist stresses under ordinary surface conditions, implying a narrow range of temperature, freedom to escape on one or more sides, and a short period of stress application. Comparatively little is known about the strength of rocks under high pressures applied on all sides (containing pressures) and high temperatures, which presumably exist deep within the earth, or under the long continued stresses of geologic periods. Thus it is that terms like *strong* and *weak*, *rigid* and *mobile*, *competent* and *incompetent*, have only relative and not absolute significance. When we say a rock is weak or mobile, we mean that it is so under a certain set of conditions or that it is so as compared with another rock. It may be relatively strong under other conditions or compared with another rock. A good deal of the discussion of diastrophism, and particularly the deeper movements within the earth, is vitiated by the use of these

[1] See appendix; also Barrell, Joseph, The strength of the earth's crust: Jour. Geol., vols. 22 and 23, 1914 and 1915.

terms, which in the nature of the case cannot be precisely defined. Their use often involves concealed assumptions as to absolute conditions, which divert the mind from the idea of relativity which they necessarily express.

Looked at broadly from a geologic standpoint, the strength of rock is due to (1) Its own inherent qualities of composition, crystallinity, and texture. (2) The amount and kind of stresses to which the rock is subjected; it has different resistance to tensional stress, to non-rotational compression, and to rotational or shearing stress. (3) The length of time during which the stress is applied and the number of times it is applied. A rock which will resist a suddenly applied stress may yield under slowly applied stress. Engineers recognize a weakening of materials under repeated stresses which they call *fatigue*, which renders the materials less resistant. (4) The temperature. Little is known about the resistance at excessively high temperatures. (5) The ease with which chemical and mineralogical changes or recrystallization may proceed. This depends both on the environment and on the nature of the rock itself. With the same stress and temperature conditions one rock may yield because these changes go on readily, while another may resist because of the absence of some factor or combination of factors which favors these processes.

With this preliminary general statement we may proceed to summarize a few quantitative facts in regard to the strength of rocks. The purpose is to present salient features and not exhaustive figures.

Elastic Strength of Rocks or Their Yielding within the Elastic Limit

Prior to crushing of most materials there is a certain amount of elastic yielding, without permanent deformation. With release of pressure the material resumes its original shape. This condition holds until the pressure passes the elastic limit.

As technically defined, *elastic limit* is that unit stress up to which the deformation is proportional to the applied load, and beyond which the deformation begins to increase in a faster ratio than the applied load. Experiments have shown that the elastic limit in rocks is extremely small. Even small loads, so far as tested, cause some permanent deformation. Bauschinger [1] found that load and deformation are not exactly proportional even at the beginning of stress application. Soft, porous rocks like sandstones show the greatest deviation from proportionality at the start. Under growing stresses they tend to approach it. Hard, compact rocks, like granite, approximate proportionality almost up to the breaking point.

Fatigue

Repeated stresses decrease the breaking strength. This is called the *fatigue* of materials, which is subject to the following laws:

"1. The rupture of a bar may be caused by repeated applications of a unit-stress less than the ultimate strength of the material.

2. The greater the range of stress, the less is the unit-stress required to produce rupture after an enormous number of applications.

3. When the unit-stress in a bar varies from 0 up to the elastic limit, an enormous number of applications is required to cause rupture.

4. A range of stress from tension into compression and back again, produces rupture with a less number of applications than the same range in stress of one kind only.

5. When the range of stress in tension is equal to that in compression, the unit-stress that produces rupture after an enormous number of applications is a little greater than one-half the elastic limit." [2]

[1] Bauschinger, J., Mittheilungen aus dem Mechanische-Technischen Laboratorium in Münich, 1875, Heft 4, p. 11.
Tests of metals: Watertown Arsenal Reports for 1894–1895.

[2] Merriman, Mansfield, Mechanics of materials: John Wiley and Sons, Inc., 11th ed., 1916, p. 353.

Quantitative values of fatigue developed in rocks during long geologic periods are naturally not subject to laboratory measurement, and are, therefore, not known.

Tensile Strength of Rocks

A rock yields by tension more easily than by compression. Under some stress conditions both compression and tension may be present (with different orientations), in which case the yielding is by tension. Put in another way, the fact that a rock yields by tension does not necessarily mean that compressional stresses are absent. The maximum and the common range of tensile strengths of rocks in pounds per square inch are about as follows: [1]

	Maximum	*Range*
Granite	1,100	500 to 800
Limestone	1,400	600 to 1,000
Dolomite	700	200 to 400
Sandstone	1,400	200 to 600

These results were obtained under ordinary laboratory conditions.

Little is known about the tensional resistance of rocks under conditions of high temperature, or under long-continued or repeated applications of stresses of the kind with which geology has often to deal.

Strength of Rocks under Bending Stresses

The transverse or bending strength of a beam of rock is closely related to tension, in that the rock first fails by tension along the convex side of the beam. However, it involves certain additional factors, in that the tension is limited to one side of the neutral surface and that compressional shear exists on the

[1] Hirschwald, J., Handbuch der bautechnischen Gesteinsprüfung: Berlin, 1912, pp. 75–82.

other. In rocks, when the elastic limit is exceeded and failure begins, the neutral surface migrates rapidly toward the concave side and practically the entire failure is by tension (see also pp. 191–192).

Crushing Strength of Rocks under Non-Rotational Compression Applied in One Direction

Two of the three principal axes of stress have a value of zero; there is no lateral support. With ordinary laboratory conditions of temperature and time the common rocks will crush by fracture when the stress reaches about the following pressures in pounds per square inch. The figures given represent the maximum and the common range of values:

	Maximum	*Range*
Granite	43,000	15,000 to 25,000
Diabase	27,000	22,000 to 26,000
Marble	23,000	9,000 to 15,000
Limestone	42,000	7,000 to 15,000
Sandstone	29,000	4,000 to 8,000
Slate	29,000	15,000 to 20,000

Impact strength. The impact strength of a rock is measured by repeatedly dropping a hammer upon the rock from known heights successively increased until the rock breaks. The weight of the hammer multiplied by the total height through which it has dropped is called the impact strength. According to Föppl,[1] the impact strength of a rock is related to its crushing strength and tenacity. The resistance to rupture of surface rocks when subjected to sudden shock, as in earthquakes, is probably closely related to the impact strength.

Below are given some of Föppl's determinations of impact

[1] Föppl, August, Die Prüfung von Steinen auf Zahigkeit mit hilfe von Schlag versuchen: Mitth. aus dem Mechanisch-technischen Laboratorium, München, Heft 30 and 32, 1912.

See also: Goldbeck, A. T., and Jackson, F. H., Physical tests of rock for road building: Office of Public Roads, Bull. 44, 1912.

strengths, converted into our standard units. The figures given for each rock represent the product of a weight in pounds and the total drop in inches which the weight must fall in order to break an inch cube of the rock. An impact strength of 2,000 indicates that a 2,000-pound weight would need to drop one inch in order to break an inch cube of the material tested. The same result would be obtained by dropping a 50-pound weight 40 inches.

Granites	3,920,	3,510,	4,570
Sandstone			217
Diabase porphyry			5,510
Graywacke			7,550
Basalt			7,890

The crushing strengths of the granites whose impact strengths are given above were 24,300, 25,200, and 32,600 pounds per square inch respectively. That of the sandstone was 12,300 pounds per square inch.

Crushing Strength of Rocks under Non-Rotational Compression where there is Lateral Support

Two of the principal stresses under the experimental conditions are equal to each other but less than the compressional stress. A considerable variety of experiments has shown that under these conditions the crushing strength of the rock is much higher than it is where there is no lateral support. Increase in the lesser principal stresses requires great increase in the major stress to cause deformation, and increase in the "stress difference" (see p. 16). Expressed in another way, the rock takes on increased strength and rigidity. Where a marble without lateral support will ordinarily crush at a pressure of 9,000 to 15,000 pounds per square inch, when sustained laterally by a steel jacket, with resisting power equivalent to about 26,000 pounds per square inch, it may require a stress difference of about 66,000 pounds to deform

it. When a disk of rock is deformed at high pressures between flat steel plates, it also takes on high rigidity even though apparently free to escape. Even a mass of soft clay so deformed becomes so rigid that it impresses itself into the steel plates. Frictional resistance may in this case supply the resistance to escape which is furnished by the steel jackets in the case above cited.

So little has been done with the actual measurements of the lesser principal stresses (the lateral support) that it is not yet possible to present a formula indicating a ratio between them and increased rigidity. Neither is it clear just what "rigidity" signifies under these conditions, i.e., the extent to which this quality is due to change in physical character of the material itself and the extent to which it is due to containing walls or friction (see pp. 324–325).

Strength of Rocks under Rotational or Shearing Stresses

(*a*) *Under ordinary surface conditions without lateral support.* The resistance to compressive shear of the common rocks usually has about the following values in pounds per square inch:

Granite	1,900 to 2,400
Limestone	1,200 to 2,000
Sandstone	1,200 to 1,800

Resistance of rocks to such stresses is, therefore, almost as small as their resistance to tension.

(*b*) *Under shearing stresses with lateral support.* This is a condition which probably exists deep within the earth. Experiments in the geologic laboratory of the University of Wisconsin, now under way, indicate that the strength or rigidity is much increased under these conditions, but that the stress difference necessary to produce deformation is far less than under non-rotational compression with similar conditions of support.

Shearing stresses exist in tension, as well as in compression, but rocks yield so much more readily by breaking normal to the tension than along inclined shearing planes, that there is no measure of their shearing strength under tension.

Crushing Strength of Rocks under Hydrostatic Compression or Equal Compression from all Directions

Bridgman [1] concludes from experimental evidence that stresses twenty times greater than the ordinary crushing strength of rocks may be necessary to close cavities in rocks by hydrostatic pressure. He tested the effect of hydrostatic pressures up to 12,000 kilograms per square centimeter (about 171,000 pounds per square inch) on hollow cylinders of quartz, barite, tourmaline, calcite, feldspar, andesite-porphyry, limestone, granite, and glass. Beyond a certain pressure, varying with the material, the walls of the bore holes flaked off. Cracks formed but no slipping took place. Granite and barite showed a slight amount of flow. Glass tubes were subjected to a hydrostatic pressure of 24,000 atmospheres without breaking. The tensile strength of the glass was only 7,000 atmospheres.[2]

Other Facts Indicating the Strength of the Earth

Available laboratory tests on the strength of rocks are not adequate to indicate the strength of the earth as a whole, but there are other methods of attack.

The resistance of the earth to tidal stresses indicates that as a whole it is stronger than steel. As stated by Reid: [3]

"If the earth yielded like a fluid, its surface would always remain at right angles to the vertical, and a pendulum would remain relatively stationary for all positions of the moon; if the earth

[1] Bridgman, P. W., Failure of cavities in crystals and rocks: Am. Jour. Sci., vol. 45, 4th ser., 1918, pp. 243–268.

[2] Bridgman, P. W., Breaking tests under hydrostatic pressure and conditions of rupture: Phil. Mag., vol. 24, 1912, p. 63.

[3] Reid, Harry Fielding, Constitution of the interior of the earth as indicated by seismological investigations: Proc. Am. Phil. Soc., vol. 54, No. 219, 1915, p. 296.

were absolutely rigid, the moon's attraction would deflect the pendulum an extremely small amount, but an amount capable of being measured."

Pendulum measurements do not give satisfactory results because the deviations are so slight; but the existence and nature of earth tides were brilliantly demonstrated by Michelson and Gale [1] by measuring the deviations in the surface of water in a pipe 500 feet long, sunk six feet in the earth. At the two ends the pipe was sealed up and the water level standing there was measured by extremely delicate methods. Variations in level were caused by the tilting of the pipe under tidal stresses. The curves of tilting show that the viscous yielding of the earth is small, as indicated by the small difference in phase between the observed and computed tides. The tides in the actual earth are .310 of what they would be if the earth were fluid. The indicated resistance of the earth to tidal stresses is greater than that of steel.

Determinations by Schweydar [2] indicate a rigidity two and one-half times that of steel. With the known low rigidity of surface rocks, he concludes that the rigidity of the interior may be ten times that of the surficial part.

The mutation of the poles is regarded as an index of high elastic rigidity.

The nature and speed of earthquake waves also indicate that the earth is rigid and elastic and not liquid, and that its elasticity increases with depth.

None of the foregoing facts, however, indicates how the earth would respond to more slowly applied and long continued stresses.

It is known that rocks have failed as far down as our observation extends, and the nature of earth movements involved in orogenic and epeirogenic disturbances or isostatic

[1] Michelson, A. A., and Gale, Henry G., The rigidity of the earth: Jour. Geol., vol. 27, 1919, pp. 585–601.

[2] Schweydar, W., On the elasticity of the earth: Naturwissenschaften (1917), Pottsdam, Germany, pt. 38.

adjustments calls for rock failure for some tens of miles down (see pp. 332–340). The possible existence of a zone of slipping or yielding immediately below a brittle crust has been one of the widely accepted postulates of geology. There is no direct proof for the existence of such a zone of weakness; the postulate involves many assumptions which may or may not be true, and lines of reasoning go back to conceptions of the origin of the earth and of the earth's diastrophism (see Chapter XIV).

If it is assumed that rocks have only the strength indicated by tests under surface conditions, it may be calculated that they are not strong enough to support a dome with the sphericity of the earth. In fact a dome of this curvature will support only one-five hundred twenty-fifth of its own weight. This is more or less irrespective of its thickness. The factor of increased rigidity with load, experimentally demonstrated, however, makes the assumption a doubtful one.

In general, then, it is established by more or less direct observation or inference from experiment that the earth is stronger than steel under rather sudden earthquake and tidal stresses, and that its rigidity increases with depth. What its strength is under more slowly continued stresses is not known, but it is probably less. There is geologic evidence that the surficial rocks have failed under stress and that this failure probably has extended downward some tens of miles. The nature and extent of failure deeper down is more or less a matter of speculation.

The apparent conflict of evidence between high rigidity and weakness may perhaps be explained by the fact that the evidence of rigidity is restricted to conditions where the stress is suddenly applied. It may be that under more slowly applied stresses, even the deeper rocks slowly yield. Also the earth as a whole may act as a rigid solid and yet have parts which yield.

The Density of Rocks and Their Isostatic Adjustment

The density of common rocks as known at the surface is indicated in the following table:

Specific gravity of common rocks.

Granite	2.63 to 2.75
Gabbro	2.9 to 3.3
Basalt	2.9 to 3.1
Rhyolite	2.4 to 2.5
Limestone	2.6 to 2.8
Dolomite	2.7 to 2.87
Shale	2.7
Slate	2.8
Sandstone	2.5 to 2.7

It is supposed that sedimentary rocks form only a thin shell, and that to a depth of ten miles fully 95 per cent of the rocks are igneous. On this assumption the average density of a ten-mile shell, with a density for each type of rock calculated as it exists at the surface, is in the neighborhood of 2.76.[1] The mean density of the earth as a whole is determined astronomically at about 5.6. If the surface rocks average 2.76 in density, it follows, therefore, that the density of part of the earth must be higher than 5, and that the density of the deep zone must be higher than that of the surface; at the center of the earth, assuming uniform gradient, it would be about 11. The actual distribution of density in the deep zone, both vertically and horizontally, is unknown.

Density can be increased by cubic compression (see pp. 306–307), and it may be supposed that part of the increase in density below the earth's surface is due to increase in pressure. But experimental evidence does not go far enough to show that the great pressures existing in the earth could sufficiently increase the density of the common rocks and minerals to account for the mean density of the earth. It has been supposed that part of the increase in density may

[1] Washington, H. S., Isostasy and rock density: Bull. Geol. Soc. Am., vol. 33, 1922, p. 402.

be accounted for by a different composition, such, for instance as the possible concentration of iron in the nucleus. Doubtfu seismologic evidence has suggested the same conclusion (se p. 275).

There is no close relation between density and strength Light rocks, like granite and quartzite, are among the stronges known rocks. Heavy rocks of high density, like basalt and limestone, are relatively weak. It is easy to think of lightness and weakness as more or less correlative, and this concealed assumption is by no means absent in the literatur relating to earth deformation.

Comparison of other planets suggests that density increase with mass, due perhaps to squeezing out of lighter materials.[1]

Areal Distribution of Density, and Isostasy[2]

Investigations covering only a small part of the earth's surface seem to indicate that the elevations are on the whole o lower density than the depressions, as if the masses were i flotational equilibrium. The crustal densities are balance against topographic relief. This condition has been calle "isostasy." The term carries an implication of hydrostati equilibrium from equality of pressure, such as flotation on liquid substratum, and of certain kinds of movements necessary to its maintenance. Its use, therefore, in describing th known facts of distribution of density to some extent confuse hypothesis with facts. It seems desirable to separate the two In this chapter we shall deal only with facts of density distribution, leaving to a later chapter (Chapter XV) consideration o what these facts imply in the way of movements. So fa as the term isostasy appears in this chapter, it refers to th facts of density distribution.

Density is determined by measurements of the intensity o

[1] Chamberlin, T. C., Diastrophism and the formative processes: Jour Geol., vol. 29, 1921, pp. 400–402.

[2] For general summary of this subject see reports by John F. Hayford an William Bowie, U. S. Coast and Geodetic Survey; also Symposium in Bull Geol. Soc. Am., vol. 33, No. 2, 1922.

gravity and by deflections of the vertical or plumb-bob. The difference between the computed and observed intensity of gravity at any point is called the *gravity anomaly*. If the higher elevations represent excess of mass, the intensity of gravity at these points should be higher than at lower points. This is not the case. When gravity anomalies are plotted it is impossible to detect from them the high and low ground. If, on the whole, the assemblage of rocks in an elevated area had the same density as the assemblage in a depressed area, the plumb-bob or vertical would tend to show a deflection toward the higher areas. It has been found in extensive geodetic surveys in this and other countries that this is not the case. While the plumb-bob is slightly deflected toward the higher areas, due to excess of mass there, the deflection is actually not as great as could be calculated on the assumption that the higher areas are of the same density as the lower areas. There is obviously some compensating pull in the lower areas, which is interpreted as an excess of density. The Coast and Geodetic Survey has determined the deflection of the plumb-bob from the astronomic vertical at some hundreds of stations in the United States. With the aid of topographic maps, the lateral pull upon the plumb-bob by topographic elevations has been calculated, without, of course, assigning any deficiency of density to the elevated areas. The calculated deflection from the vertical, under the influence of the topography, has been found in each case to be much larger than the actually observed deflection, though usually in the same direction. The obvious inference is that there is a counteracting pull downward due to excess of density at the point of observation, in other words, that there is excess of density in the topographic depressions corresponding to deficiencies in the elevations. There is thus a certain balance or *compensation* between density and topography — areas of low density being compensated by higher elevation and heavy areas by depression. The following quotations are from Hayford:[1]

[1] Hayford, John F., The figure of the earth and isostasy from measurements in the United States: U. S. Coast and Geodetic Survey, 1909, pp. 65, 166, and 169.

"The logical conclusion from the study of the geoid contours for the United States, taken in connection with the fact already noted that the computed topographic deflections are much larger than the observed deflections of the vertical, is that some influence must be in operation which produces an incomplete counterbalancing of the deflections produced by the topography, leaving much smaller deflections in the same direction."

* * * *

"The writer believes that the stress-differences in and about the United States have been so reduced by the isostatic compensation that they are less than one-twentieth as great as they would be if the continent were maintained in its elevated position and the ocean floor maintained in its depressed position by the rigidity of the earth."

* * * *

"It is certain, from the results of this investigation, that the continent as a whole is closely compensated and that areas as large as States are also compensated. It is the writer's belief that each area as large as one degree square is generally largely compensated. The writer predicts that future investigations will show that the maximum horizontal extent which a topographic feature may have and still escape compensation is between one square mile and one square degree. This prediction is based, in part, upon a consideration of the mechanics of the problem."

Regional Distribution of Compensation

It is not easy to determine how far each small area or column may be compensated independently of surrounding regions; in other words, how much area should be taken into account in figuring whether there is or is not compensation. From an analysis of data made by the United States Coast and Geodetic Survey, Bowie [1] concludes:

"It was found that the gravity anomalies were reduced as effectively by regional distribution of compensation out to a distance of 36 miles from the station as by local compensation; but when the compensation was distributed to a distance of 100 miles from the

[1] Bowie, William, Theory of isostasy — a geological problem: Bull. Geol. Soc. of Am., vol. 33, 1922, p. 279.

station, the gravity anomalies became systematic in their appearance and it was found that there was a definite relation between the elevation of the station and the gravity anomaly. It is reasonably certain that there is no strictly local distribution of the isostatic compensation, but it is equally improbable that the compensation is distributed for great distances from the topographic features. After consideration of the resistance to vertical movement by a column of the isostatic shell, the speaker believes that the column which may be in isostatic equilibrium, independently of the surrounding areas, is of the order of magnitude of one square degree at the equator, or about 70 miles square."

Regional Distribution of Densities in Relation to Kinds of Rock and Age of Formation

A generalization which has been widely current among geologists for some time is that mountains and continents are on the whole composed of more acidic rocks than are ocean basins, but the detailed evidence has been fragmentary. Washington [1] has recently shown for North America that the actual determined and computed specific gravities of the rocks in higher areas are lower than they are in the lower areas; that "the average density of the igneous rocks of a region varies in the opposite sense as the average altitude," that the higher regions are dominantly acidic, and the lower regions are dominantly basic. Co-magmatic regions show striking correspondence to areal distribution of densities. He reaches this conclusion by two methods (1) by correlation of average density determinations, and (2) by starting with the average composition of the rocks of an area, reducing this to mineral composition, according to a uniform or normative method, and from the mineral composition calculating the average density, the density of the several constituent minerals being known.

In this calculation the sedimentary rocks are omitted, on the ground that they are insignificant in amount as compared with igneous rocks, to the depths involved in geodetic calculations.

[1] Washington, H. S., Isostasy and rock density: Bull. Geol. Soc. Am., vol. 33, 1922, p. 393.

It has been particularly noted that granites, which are light rocks, are very characteristic of the higher points of many mountain chains. Nevertheless, equally great areas of granite exist in low, peneplained areas of pre-Cambrian and later periods in many parts of the world. The immense areas of granite in the Laurentian peneplain of Canada may be cited as an illustration.

It would seem to follow that whatever localizes molten rock also determines the distribution of density. Thus, a granite batholith invading an area previously occupied by gabbro or basalt substitutes its own density for that of the basalt, involving a reduction of density of about 13 per cent.

There appears to be a definite relation in density between the pre-Cambrian and the Cenozoic formations, according to the results of geodetic surveys in United States, Canada, and India. The pre-Cambrian areas have predominantly an excess of density and the Cenozoic areas are light, i.e., show deficiency of density.[1]

Recent delta deposits have been especially measured to see if these accumulating loads indicate any excess of mass or any departure from isostatic adjustment, and it has been found that they are actually deficient in density.

Depth of Isostatic Compensation

The depth through which the areal differences of density are supposed to extend has been called the depth of compensation. A plane at this depth would support an equal weight of material above on each unit of area, the density of the material being compensated by the height of the column; below this depth, the density is supposed to be uniform. Postulating the existence of such a plane of complete compensation, Hayford assumed various arbitrary depths in order to find out which one corresponded most closely to the facts of the gravity observations. For each of these arbitrary depths, three alternative distributions of density were assumed — (1) uniform vertical distribution of density to the depth of complete compensation;

[1] Bowie, W., *loc. cit.*, p. 275.

(2) a gradually diminishing difference in density to this depth; (3) a maximum difference in density at some intermediate point. Depending on distribution of density chosen, the depth of complete compensation was found to be between 60 and 150 miles. With a uniform distribution of density a depth of compensation of 76 miles was found best to correspond with the plumb-bob observations.

Later estimates by Bowie are about 37 and 60 miles on the assumption of equal vertical distribution of density.

In the calculations of the geodesists, a definite relationship exists between depth of compensation, topographic relief, and variation in rock density. Each different value assumed for the depth of compensation requires the assumption of a definite variation in the density of the earth's crust. For two rock columns differing in relief 10,000 feet, the difference in density for a depth of compensation of 60 miles is 3.3%, and for a depth of compensation of 120 miles it is 1.7%. Now it is known that the surface density of the lithosphere may vary by 10% or more. This variation is not used by geodesists in their calculations of the depth of compensation, as they derive the variation in density from the depth of compensation chosen. If the actual variation in rock density were taken into account, this would obviously affect the result.

Washington [1] calculates a depth of 37 miles by comparing the normative densities of continents and ocean floors. He calls this the *isopiestic* level.

The actual distribution of density with depth is not known. All that can be said is that *if* the differences in assumed density at the surface are to be maintained uniformly to the depths noted, this would explain and account for the observed differences in elevation at the surface, as well or better than any other assumed depth. There is no reason why the depth of compensation should not, with different assumptions as to vertical distribution of density, be uniformly greater or less, or that it should not be highly irregular in depth. The hori-

[1] Washington, H. S., Isostasy and rock density: Bull. Geol. Soc. Am., vol. 33, 1922, p. 402.

zontal distribution of densities is highly irregular; why not the vertical distribution likewise? A uniform depth of compensation is essentially a mathematical convention for the purpose of easy reference, which lessens the number of variables to be handled in computation. There is no proof that it corresponds to any physical reality; it is little more than a guess. Geodesists differ in the assumptions used, but from their constant handling of the data have come to show a strong leaning toward the assumption of a uniform depth of compensation as a working hypothesis.

As a matter of fact the existence of a uniform depth of compensation is not essential to the theory of isostasy.

Comment on Observed Variations from Perfect Isostatic Adjustment

It is to be noted that the investigations of isostasy are based on geodetic surveys confined largely to the areas of the United States, Canada, and India, which comprise only 10 or 12 per cent of the land surface of the globe. The ocean areas, where no data exist, cover 70 per cent of the globe. Therefore, the geodetic investigations in isostasy represent only about 3 per cent of the entire surface of the earth. Miscellaneous pendulum determinations of gravity have been made over the oceans, the results being for the most part normal, but adequate gravity determinations over the oceans have not yet been made.

Within the area studied, the average variations from perfect isostatic adjustment, based on assumption of uniform depth of compensation, according to Hayford, are equivalent to a load of 250 feet of rock. Locally there are many wider variations.

Geodesists make particular note of the fact that Cenozoic sedimentary deposits, and particularly deltas and other recent shore deposits, are deficient in density as compared with the requirements of perfect isostasy. These areas have recently accumulated heavy loads of sediments and should show an excess of density over isostatic requirements. The deficiency

is ascribed to the fact that the sediments are largely unconsolidated and have low density, which abnormally affects observations made at nearby stations. It is argued that an observed deficiency is probably compensated by heavier material in the columns below, thus bringing the columns as a whole into isostatic adjustment. This is put on the ground that heavy sedimentation must necessarily increase the load, and would not be likely to cause deficiency in mass for the column as a whole. This argument obviously presupposes the correctness of the hypothesis that the earth actually possesses perfect isostatic adjustment, and is not inductive reasoning from the observed facts.

Likewise, it is found that the pre-Cambrian areas on the whole show an excess of density over the requirements of perfect isostasy. Again the explanation is that there is probable compensation by lighter material below. While geologists are ready to conceive the possibility, or even the probability, that there are denser materials below areas of light, unconsolidated sediments, they do not know any facts which would favor the belief that there are lighter materials beneath the dense basement rocks of the pre-Cambrian.

Areas of erosion, like the pre-Cambrian and others, might be supposed to show deficiency of density to the extent to which isostatic adjustment has lagged behind the unloading. To a large extent they show actual excess of density. This fact is construed by geodesists as meaning that there has been complete or excess compensation, and that the density of rocks near the surface unduly affects the observations.

Notwithstanding the above doubts and qualifications, there remains a notable degree of concordance of geodetic observations with the requirements of a high degree of isostatic adjustment.

When we consider the distributions of densities, vertically and laterally, which are involved in isostatic adjustment, the information is much less decisive, as indicated under preceding headings.

For these reasons the geologist finds it extraordinarily difficult to form a definite idea of the extent to which isostatic balance actually exists and of the distribution of masses of varying density, though he cannot but accept the geodesist's conclusion that some kind of isostatic balance is a reality. The geologist has considerable doubts as to the mechanics of the movements postulated by the theory of isostasy, discussed in a succeeding chapter.

The Volume Changes in Rocks

Volume Changes Due to Temperature Variation

The coefficients of linear expansion of a few of the common minerals and rocks for a rise of one degree Centigrade are reported as follows: [1]

Rock.	*Temperature range.*	*Coefficient of linear expansion.*
Calcite (along principal axis)	0 –100° mean	.000026315
" (perpendicular)	0 –100° mean	.000005440
Quartz	19°– 46°	.0000119
"	50°– 60°	.00001177
" (parallel to principal axis)	0 –100° mean	.000007971
" (perpendicular to axis)	0 –100° mean	.000013371
White marble	15°–100° C.	.00000117
White Carrara	0 –100°	.00000849
Black marble	0 –100° C.	.00000445
Limestone	0 –100° C.	.00000251
Granite	0 –100° C.	.00000868

The results of different investigators vary somewhat, and the same samples show different coefficients of expansion for different intervals on the temperature scale.

The coefficient of cubical expansion of a substance is the ratio between the increase in volume when its temperature rises one degree Centigrade and the original volume. It is about three times as great as the coefficient of linear expansion.

If the coefficient of linear expansion of granite is taken as .00000868, then its coefficient of cubical expansion is about .00002604. The volume of one cubic mile of granite after

[1] Castell-Evans, J., Physico-Chemical Tables, London, 1902; also Landolt and Börnstein, Meyerhoffer Tabbelen, 1905.

its temperature is raised by one degree Centigrade is 1.00002604 cubic miles. After a rise of 1000 degrees Centigrade its volume would be 1.02604; the percentage of volume increase is 2.6 per cent.

Data compiled by Daly [1] indicate that the volume increase of granite and gabbro when heated from 20° to 1,300° C. would be between 3 and 4 per cent. This compilation is based on experimental work by Douglas.[2]

At higher temperatures most rocks undergo volume changes due to chemical reactions, and certain minerals pass through conversion points at which their crystallization changes. The rock minerals which form on the higher temperature side of a conversion point are less dense than those existing below the conversion point. Quartz in its change to tridymite at 800° C. illustrates this principle; the density decrease in this case is about 13 per cent.

The increase in volume when rocks change from the solid to the liquid state at ordinary pressures is considerable. Barus [3] states that when diabase changes to glass the volume increase is about 10 per cent, and that the change from liquid diabase to glass involves a contraction of 3.5 per cent to 4 per cent. Daly's data show the volume increase of gabbro at 1,000° C., changed to the liquid form having the same temperature, to be 7.65 per cent. Volume change for granite under the same condition is 10.7 per cent.

Within the molten state volume changes due to variations in temperature are more marked in general than in the solid state. From data cited by Daly the volume increase in liquid gabbro for a rise of temperature from 1,000° to 1,300° C. is .46 per cent and for granite .74 per cent.

The preceding estimates do not take pressure into account.

[1] Daly, R. A., Igneous rocks and their origin: McGraw-Hill Book Co., New York, 1914, p. 202.

[2] Douglas, J. A., On changes of physical constants which take place in certain minerals and igneous rocks on the passage from the crystalline to the glassy state: Quart. Jour. Geol. Soc., vol. 63, 1907, p. 145.

[3] Barus, Carl, High temperature work in igneous fusion and ebullition, chiefly in relation to pressure: Bull. 103, U. S. Geol. Survey, 1893, pp. 25–26.

Tammann [1] concluded that the volume decrease involved in the crystallization of minerals from magmas becomes less at higher pressures and finally is reversed, but Bridgman's [2] later studies of the melting curves of twenty substances found no evidence for this reversal.

When a magma crystallizes there is release of volatile constituents, especially water, with rapid increase of pressure; many times the original vapor pressure of the magma may result from the crystallization of but a small proportion of the non-volatile materials. This release may be catastrophic in violence under conditions which allow accumulation of pressure.[3]

Volume Changes in Rocks Due to Compression

Elastic compressibility of rocks (experimental). Experiments in the Geophysical Laboratory at Washington [4] indicate the following compressibilities, per megabar,[5] expressed in parts per 10,000,000 at a pressure of 2,000 megabars:

Marble	1.41
Granite	2.13
Basalt	1.88
Diabase	1.26

At high pressures, compressibility is less than at low pressures. The compressibility per megabar of the preceding substances, when under a pressure of 10,000 megabars, expressed in parts per 10,000,000 is as follows:

Marble	1.41
Granite	1.84
Basalt	1.55
Diabase	1.26

[1] Tammann, G., Kristallisieren und Schmelzen, 1903.

[2] Bridgman, P. W., Change of phase under pressure: Physical Review, vol. 3, 1914–1915.

[3] Morey, George W., The development of pressure in magmas as a result of crystallization: Jour. Wash. Acad. Sci., vol. 12, 1922, pp. 219–230.

[4] Williamson, E. D., Change of the physical properties of materials with pressure: Geophysical Lab., Wash., Pub. No. 446, 1922.

[5] A megabar is equivalent to 1,000,000 dynes per square centimeter or 987 atmospheres.

For a cubic compression of 200,000 pounds per square inch, or about 13,500 megabars, the percentage volume decrease would be as follows:

Granite	.248	per cent
Marble	.190	" "
Basalt	.209	" "
Diabase	.170	" "

The compressibility of diamond and quartz per megabar at 0 megabars is respectively .18 and 2.7 parts in 10,000,000. Diamond is the least compressible substance known. Its nearest competitor, tungsten, is nearly twice as compressible and the majority of solids are more than ten times as compressible.

Permanent compressibility of rocks (experimental). Adams [1] found no increase in density in marbles which had been subjected to high differential pressures on all sides.

Experimental results on the permanent compressibility of metals seem to be at variance.

Bridgman [2] subjected metals to a dydrostatic compression of 25,000 to 30,000 atmospheres without affecting changes in density.

Kahlbaum [3] and Lea and Thomas [4] report a decrease in density for metals subjected to unequal non-rotational stress. Kahlbaum believed that he used hydrostatic compression but this view is held to be erroneous by Bridgman.

Volume Changes Due to Breaking

The foregoing experiments on volume change cover only changes in solid rocks without rupture. In large masses in the

[1] Adams, F. D., and Coker, E. G., An experimental investigation into the flow of rocks — The flow of marble: Am. Jour. Sci., vol. 29, 4th ser., 1910, p. 465.

[2] Bridgman, P. W., Breaking tests under hydrostatic pressure and conditions of rupture: Phil. Mag., vol. 24, 1912, p. 63.

[3] Kahlbaum, W. A., Roth, K., and Siedler, P., Ueber Metall. distillation und über distillierte Metalle; Zeitschr. Anor. Chem., vol. 29, 1902, p. 254.

[4] Lea, F. C., and Thomas, W. N., Change in density of mild steel strained by compression beyond the yield point: Eng:, vol. 100, 1915, pp. 1–3.

earth it is necessary also to consider the increase of volume due to breaking and separation of the parts. In a breccia the volume increase may be 50 per cent. In joints and faults it is less, but still considerable; no accurate estimate of average conditions is possible. Becker [1] has called attention to the quantitative adequacy of this factor to explain mountain uplifts and their relations to isostasy.

Volume Changes Due to Metamorphism

Weathering or katamorphic alterations ultimately change igneous rocks into materials which when distributed form sediments and salts of the sea. The increase in volume in the change from average igneous rock to sediments, including the solid equivalent of the salts of the sea and development of pore space, is about 37 per cent. For the sediments alone, including pore space, the volume increase is at least 28 per cent, and without the pore space, 11 per cent of the original rock.

Katamorphism increases the volume: (*a*) by mechanically distintegrating and separating materials, forming openings; (*b*) by the addition of water, carbon dioxide, and oxygen; and (*c*) by decreasing the average specific gravity of the minerals. This does not mean that minerals of higher specific gravity may not be formed, as for instance limonite, but the average density is lowered when all the resulting minerals are taken into account.

It is of interest to note that the principal substances added, — oxygen, hydrogen, and carbon, — are all of light weight. It is to be expected, therefore, that the molecular combinations into which they enter should be on the whole lightened and made less dense.

Cementation of sediments involves decrease in volume in so far as there is settling or slumping. On the other hand, the introduction of cements, which crystallize from solution,

[1] Becker, G. F., Note on mean density of fractured rocks: Jour. Wash. Acad. Sci., vol. 4, 1914, pp. 429–431.

might have the effect in some cases of actually increasing volume, due to the crystallizing power of the cementing minerals.

During anamorphism of sediments by rock flowage there is a diminution in volume due to reduction of pore space, due to the elimination of lighter constituents, like carbon dioxide, water, oxygen, and possibly other substances, and due, in some cases, to the development of minerals of higher density. It is true that the molecular volumes of certain minerals developed by anamorphism are greater than the molecular volumes of the replaced primary minerals. For example, the molecular volume of wollastonite is greater than that of calcite, but wollastonite should be compared with calcite plus quartz, in which case a diminution in volume appears. Notwithstanding the fact that some of the minerals developed under anamorphism are of larger specific volume than the primary minerals from which they are derived, it is true that the characteristic schist-making minerals are, on the whole, of higher specific gravity than those characteristic of katamorphism. In other words, proportions of hornblende, biotite, muscovite, and chlorite common in schists give distinctly higher specific gravity to the rock than a combination of kaolin, quartz, and calcite. When the decrease in porosity and the elimination of certain substances not needed for the schist-making minerals are also considered, a decrease in volume becomes self-evident.

Under contact metamorphism the volume may be likewise reduced, but where there is considerable addition of substances from the igneous rock, metasomatically replacing constituents of the sediment, there may be little or no change in volume.

The volume changes above summarized have not been fully measured, but they are known to be considerable. For instance, the change of mud or clay to shale involves a volume reduction of not less than 17 per cent, and from shale to slate the reduction is about 12 per cent.

It is apparent from the foregoing that compressibility and

volume changes due to temperature change are small as compared with volume changes resulting from rock fracture and metamorphism.

Other Changes of the Physical Properties of Materials with Pressure (Experimental) [1]

Increase of pressure increases electrical conductivity. So definite is this that in special cases the increase in conductivity can be used as a measure of the pressure.

Possibly also other physical properties, such as heat conductivity, may be affected. Unequal compression lowers the melting point of certain substances. Cubic or uniform pressure raises the melting point.

Pressure also has a profound effect on chemical and mineral changes, diffusion, viscosity, and other properties, for which the study of metamorphism furnishes abundant evidence.

Here is a great and promising field for investigation. In interpreting physical conditions deep below the surface we have had to rely in the past mainly on the physical properties of the rocks as known under ordinary surface conditions. If the properties deep below the surface are markedly different, as suggested by experimental work and inferred from metamorphic changes, many of our elementary inferences as to the behavior of rocks under conditions of high pressure beneath the surface may have to be modified.

General Comments and Summary

Our knowledge of the behavior of rocks under pressure is limited to a comparatively narrow range of conditions attainable in the laboratory. Experiments are limited to a range far short of that necessary to reproduce conditions deep within the earth. They cover a range representing only a zone extending fifty kilometers below the surface. For ordinary

[1] See various reports of the Geophysical Laboratory, Washington.

temperatures exact measurements can be made up to 200,000 pounds per square inch, and under special conditions to higher pressures. At higher temperatures it is not possible to use such high pressures. At the highest temperature used by the Geophysical Laboratory of Washington, 1300° C., it has been possible to use only one-tenth of the pressure which could be used at room temperature.

The results obtained within these limits do not tell exactly how rocks behave under greater pressures deep within the earth, although they do yield valuable suggestions which complement information as to density and rigidity furnished by astronomical, tidal, and seismic investigation.

Even within the range of experiment there are still large gaps in our information, particularly with reference to shearing strength under high pressures. Still further, much remains to be done in the way of correlation and interpretation of available information.

A brief generalization of the known facts relating to the strength, density, and volume changes within the earth is attempted below.

The compressive strength of crystalline rocks under non-rotational stresses is thirty to forty times their tensile strength. The ratio varies for different kinds of rocks and for rocks in differing states of consolidation. In general, igneous rocks have a higher ratio of tensile to compressive strength than sedimentary rocks.

The strength of rocks under shearing stresses is far less than under non-rotational compression, being not far above the tensile strength.

Where there is lateral support the strength of the rocks is greatly increased under both non-rotational and shearing stresses, but less stress difference is required for deformation by shearing than by non-rotational compression. Inferences as to strength of rocks far beneath the surface have been based for the most part on the assumption that the stresses are non-rotational, and little attention has been paid to the

equal or greater probability that they are rotational, in which case the strength of the rocks is very much less.

Under sudden stresses the earth is a solid, more elastic and rigid than steel.

The rigidity is far greater than that of surface rocks. Experimental tests indicate that rigidity is induced by high pressure.

On the other hand, there is abundant evidence that the earth has locally failed in the zone of observation by fracture, flowage, and vulcanism, and presumably for some distance at least below our depth of observation.

The earth as a whole has a mean density far greater than that of surface rocks, indicating that the interior of the earth is much denser than the surface rocks. It is not known whether this greater interior density is due to compression or to the existence there of larger proportions of dense substances like iron. Judging from the compressibility of rocks within experimental range, the high density of the interior can not be ascribed to compression alone.

The inequalities of the earth's surface are more or less balanced against differences in density, the elevations being light and the depressions heavy.

Changes in volume and density of rock masses are known to be accomplished by compression without rupture, by crushing, by temperature changes, and by the processes of metamorphism. Of these changes, the ones due to compressibility and to temperature changes are quantitatively less than the others. Considerable change in distribution of density within the earth is due to migrations of molten rock.

CHAPTER XIV

GENERAL CONSIDERATION OF THE STRUCTURAL CONDITIONS IN THE INTERIOR OF THE EARTH

Is the interior of the earth solid or liquid, crystalline or amorphous? Is it composed of materials different from those of the surface? What are the distribution and nature of its movements? Does it contain a zone of rock flowage or an asthenosphere? These questions have occupied the attention of geologists since the birth of the science. Final answers are not yet possible and perhaps they never will be. There are, however, a few fairly well established facts which must be taken into account by any hypothesis.

The earth's interior is more rigid than steel under sudden stresses like earthquakes or tidal pulls. The viscous yielding under these stresses is negligible.[1] The earth transmits earthquake waves like a solid. It follows that the rigidity of the interior is much greater than that of the surficial rocks, which are less rigid than steel. Experimental work also points to greater rigidity with depth. The interior is over twice as dense as the surficial rocks. The temperatures are so high that any of its materials would be liquid were it not for the

[1] Chamberlin concludes:

"1. That the earth-body is essentially an elastico-rigid spheroid in which the molten and viscous elements are so far subordinate that the larger problems of the earth-body are to be solved on the elastico-rigid basis.

2. That the method of yield of the elastico-rigid body is dominantly *idioatomic*, or *idiomolecular*, that is, takes the form of progressive reorganization atom by atom or molecule by molecule, each acting individually and successively rather than collectively and simultaneously as in fluid or viscous bodies."

* * * *

"When the elastic limit of deep-lying matter is reached and a new state is inevitable, this new state is most commonly *a new elastico-rigid state*, so assumed as to relieve the enforcing stress. This is particularly the case when the balanced pressures are high and the differential stresses come slowly into action." Chamberlin, T. C., Study of fundamental problems of geology: Year Book No. 21, Carnegie Inst., 1922, pp. 359 and 360.

fact that the pressures are also tremendous. In the range of our experimental observation, liquefaction usually requires increase in volume and sufficient pressure can prevent increase in volume, and, therefore, liquefaction. Both pressure and temperature increase to the center of the earth, but the rate of increase is not known for either. Various assumptions are possible, under some of which the pressures are uniformly too high to allow liquefaction and under others of which liquefaction seems possible. The prevailing view is that the pressures are on the whole sufficient to prevent liquefaction, but that locally, and particularly near the surface, diastrophic movements may sufficiently relieve the pressure to allow earth materials to liquefy, thus explaining the origin of vulcanism. The exact physical condition of a substance so highly heated that it would instantly liquefy if under surface conditions, and yet under such high pressures that it remains a solid, is not known.

The increase in density with depth has been ascribed by some investigators to increased pressure, perhaps requiring new molecular combinations, but no great change in the proportion of elements as compared with those of the surficial rocks. The fact is emphasized that earthquake waves show that rigidity increases with depth faster than density. Others regard pressure as insufficient to produce these high densities, and ascribe them to the existence of a metallic core. The requirements are met by the assumption of a core of metallic iron, or iron and nickel, of varying dimensions, depending on the hypothesis used.

The evidence from earthquakes indicates not only a solid earth but the possible existence of a more rigid and dense core, with some degree of homogeneity. It also indicates the possibility of a change in condition immediately below a comparatively thin surficial shell (see pp. 275–276). None of these conclusions as to zonal arrangement is finally established.

With these salient considerations in mind, there is still room for wide difference of opinion as to how the condition of the

interior shall be described. Under sudden stress it is certainly solid, elastic, and rigid, but it may also be plastic and even partake of the properties of a liquid under long-continued stress. In a large way it acts as a homogeneous body, but this homogeneity may be the sum or result of minor heterogeneities. The fact that it is elastic and rigid under sudden stress suggests that it may be made up of crystalline, heterogeneous rocks, similar to the rocks of the surface, but the possibility is not precluded that the materials may be amorphous or isotropic. The crust to a possible depth of ten miles has not acted as a homogeneous structural unit, but has yielded by both fracture and flow in all possible directions. It has shown its capacity to adapt itself to any combination of vertical or horizontal stresses, whether these be non-rotational or rotational. The yielding has shown almost no tendency toward concentration in simple, easily defined zones, either vertical, horizontal, or inclined.

That there are movements in the unseen below seems to be clearly indicated by the evidence of mountain-building and other crustal deformation (see pp. 241–243). It is commonly assumed that the movements are confined to a single zone of mobility or weakness. The conception of such a zone fits the hypothesis of movement involved in isostasy, the hypothesis of a rigid shell of the earth adjusting itself to a shrinking interior, and the apparent requirements of mountain-building, which seem to involve a slipping of surficial strata over deep strata. The existence of such a zone has also been inferred from the assumption that rock flowage must occur under such conditions of temperature and pressure as are supposed to obtain at great depth.

As to depth and thickness of such a hypothetical zone of weakness, or flowage, or mobility, there has been little agreement.

Van Hise [1] assigned a depth of only six miles to the top of

[1] Van Hise, C. R., Principles of North American pre-Cambrian geology: 16th Ann. Rept., U. S. Geol. Survey, pt. 1, 1896, p. 593.

the zone of rock flowage, though with the important reservation that increased rigidity under containing pressures would greatly increase this figure.

Adams and Bancroft,[1] on the basis of experiments with rock failure under great containing pressures, conclude that the amount of tangential thrust required to produce movements increases so rapidly below the surface "that the great movements of adjustment by rock flow or transference of material in the earth's crust from one point to another — other than the transference of rock in a molten condition — must take place comparatively near the surface," and that the ease of movement "increases rapidly in proportion to their nearness to the surface." The mobile zone thus implied is inferred from experimental results to be limited to depths within 35 miles, below which a condition of no mobility seems to be assumed.

Gilbert[2] conceived "a relatively mobile layer separating a less mobile layer above from a nearly immobile nucleus," the mobile layer agreeing in depth with the depth of isostatic compensation.

Barrell[3] called this weak zone the *asthenosphere* and placed its provisional boundaries at depths of 75 and 450 miles from the surface. This he conceived to underlie the depth of isostatic compensation.

Hayford[4] assumed concentration of movement within the lower part of the zone of isostatic compensation, that is, within 75 miles of the surface. Bowie[4] and others place it below the depth of compensation.

Willis[5] concludes that there is a zone of adjustment below

[1] Adams, Frank D., and Bancroft, J. Austen, On the amount of internal friction developed in rocks during deformation and on the relative plasticity of different types of rocks: Jour. Geol., vol. 25, 1917, p. 635.

[2] Gilbert, G. K., Interpretation of anomalies of gravity: Prof. Paper 85-C, U. S. Geol. Survey, 1913, pp. 34-35.

[3] Barrell, Joseph, The strength of the earth's crust: Jour. Geol., vol. 22, 1914, p. 680.

[4] *Loc. cit.*

[5] Willis, Bailey, Discoidal structure of the lithosphere: Bull. Geol. Soc. Am., vol. 31, 1920, p. 274.

40 miles and extending to the base of the asthenosphere, and that the adjustments necessary to isostatic undertow take place mainly between 45 and 100 miles from the surface.

In contrast to this conception of the existence of a single zone concentric to the earth, is the conception of T. C. Chamberlin and R. T. Chamberlin, based on R. T. Chamberlin's study of mountain deformation, that movements may be on intersecting planes inclined to the surface, bounding wedge-shaped masses. "No undertow in a hypothetical mobile substratum is necessarily involved and none is postulated."[1]

These are only a few of the views that might be cited to indicate the wide range of hypotheses possible as to depth, number, and attitude of deep mobile zones. The very diversity of these views emphasizes the restricted range of known facts. The requirement of proof naturally rests most heavily on hypotheses which most precisely restrict the locus of movement. So many assumptions must enter into this proof that in our present state of knowledge it can not be rigorous. The safest scientific attitude for the time being would seem to be one of rigid adherence to the known facts, and the recognition of the possibility of more than one hypothesis to explain them. This is not incompatible with a sympathetic attitude toward the efforts of those attempting proof of a single hypothesis.

Until the time comes when it is possible to furnish definite proof of any specific localization of movement, our own inclination is to keep clearly in mind the irregular distribution of movements within the zone of observation, already summarized, as perhaps the best guide to the condition that may be assumed at least for some distance below our lowest observations. This measuring stick is short, but there are some reasons for believing that it is as good as any yet available to measure our course through the complex of hypotheses possible in the deep zone. Especially is it desirable to keep in mind the fact that cleavage, indicating rock flowage, as

[1] Chamberlin, T. C., Diastrophism and the formative processes: Jour. Geol., vol. 26, 1918, p. 197.

observed in the deepest part of our zone of observation, does not in general have an attitude required by the conception of tangential shearing in a horizontal mobile zone. Much of it has an attitude suggesting shearing in vertical or inclined mobile zones. This does not disprove a different attitude below, but it does eliminate an affirmative bearing on the question which has sometimes been implied.

Nature of deep movements. Whatever the distribution of deep movements, it remains to consider the manner or processes through which they are accomplished — whether by plastic flow, as usually assumed, by fracture, or by some combination of these kinds of deformation. The deformed rocks have not been seen nor have the environmental conditions been accurately measured; yet there are weighty considerations favoring the view that movements may be by rock flowage.

At the outset it should be understood that *mobility, rock flowage, weakness,* and *incompetence* are not synonymous terms, though they have been used more or less interchangeably in geologic literature. It is quite conceivable that the concentration of forces in a zone may cause mobility, even though the rocks are inherently no weaker than the rocks in other zones adjacent. Mobility does not necessarily mean rock flowage, for it may also be expressed by rock fracture or vulcanism. Rock flowage may or may not imply weakness. Rock flowage is usually assumed to imply weakness as compared with rock fracture. Both are evidence of failure, but the energy factors are not well enough known to warrant the conclusion that a rock which yields by flowage is weaker than a rock which yields by fracture, when all the environmental conditions are taken into account. It may well be that rocks undergoing rock flowage are inherently very strong, and that their yielding is due to concentration of environmental forces sufficient to overcome this strength, — as is suggested by the great increase of rigidity, requiring increase of stress differences to overcome, found in experimental reproduction of rock flowage.

The common conditions causing rock fracture near the earth's surface and in the laboratory are known to most students. It is merely a matter of applying pressure greater than the crushing strength to rocks free to expand toward one or more free sides.

When asked under what conditions rocks will flow the student's answer is frequently "more pressure." Mere increase of pressure under the ordinary conditions cited means more and quicker fracture. And yet the answer contains part of the truth, for under certain other conditions more pressure will accomplish rock flowage. The full answer requires consideration of the inherent weakness of the rock itself, its susceptibility to the chemical and mineralogic changes which aid rock flowage, the time element or the speed with which stresses are applied, and the nature of the pressure.

Weak rocks, like unconsolidated mud, clay, sand, or marl, may flow under pressure, even when this amounts to no more than their own weight.

Examination of rocks which have undergone rock flowage discloses the fact that these have undergone extensive chemical and mineralogical changes, often referred to generally as recrystallization, and that these changes in fact have been the principal means of adjustment of the materials to the new forms required by the deformative stresses. Certain minerals and rocks are more susceptible to these changes than others, and therefore more readily undergo rock flowage. For instance, the calcite of limestone is readily dissolved and recrystallized. After extensive deformation, the microscope may show few interior fractures. The crystals are large, clear, and limpid, indicating that healing and constructive processes of recrystallization have been at work. Muds and clays are susceptible to these processes through the combination of clay with the bases present, producing silicates (principally the micas). A rock made of clear sand is less susceptible to the processes of recrystallization, although they are effective to some extent.

The ease and speed of recrystallization are also determined to some extent by the nature of solutions permeating the rock and by the temperature. Highly mineralized hot solutions may effect rapid recrystallization of the essential materials of the rock, thereby favoring rock flowage. This is evidenced by flowage of hard rocks at moderate depths at batholithic contacts. Also, facts of physical chemistry show that increase of temperature increases molecular activity, hastens endothermic reactions (anamorphic reactions are largely endothermic), increases solution, both liquid and solid, and hence recrystallization, and decreases viscosity or internal friction.

The time element obviously plays a considerable rôle. A suddenly applied force may cause a comparatively soft material to fracture, while the slow application of force may cause it to flow. Marble gravestones suspended by the ends for many years are said to have sagged without rupture. Rocks may be subjected to moderate forces for immense periods of time, and under these conditions it is entirely possible that they may flow, even though they would fracture in the laboratory when subjected to the same forces for only a limited time.

Even the hardest rocks have been made to flow in the laboratory, without loss of strength, by proper application of forces. The essential requirement seems to be that the rock be held on all sides under containing pressures, and that the principal deforming pressure be largely in excess of the crushing strength of the rock. These experiments are summarized in the appendix.

Under the microscope, the deformation under these conditions is seen to have been accomplished by granulation and slicing of the particles, sometimes along more or less definite planes or zones of shear, and by the gliding of calcite crystals along twinning planes. The particles are still held together, but it is not flowage of the kind usually seen in nature where there is an extensive development of new and characteristic minerals. It is a structure which really falls in the intermediate field between fracture and flowage. Possibly with

longer time and proper conditions of temperature and mineralizers, the kind of flowage observed in nature could be duplicated.

Under these experimental conditions it is found that a very large stress difference (meaning the difference between the greatest and least of the principal stresses acting upon the rock, see pp. 290–291), is necessary to cause rock flowage. The fact that the material is under containing pressure imparts a rigidity to it, the overcoming of which requires greater stress difference than would be necessary in fracturing a rock not so held in by containers. Under a pressure of 12,000 atmospheres ordinary paraffin wax becomes so hard that it can deform steel. Soft rubber may be made harder than mild steel. A substance like a dime or a brass tack, when imbedded in steel and then subjected to enormous pressure, acquires rigidity so great that it will impress itself on the steel before flowing. A disk of clay deformed between two flat plates, even without restraining walls on the sides, takes on increased rigidity, perhaps because of frictional resistance which hinders lateral flow.

Adams[1] and Pfaff[2] also found in their experiments that when rocks were under pressure enormously greater than their ordinary crushing strength, they would not flow through a small hole bored in the side of the steel jacket, nor would small holes in the rock become closed; and it was concluded that a high degree of artificial rigidity had been induced in the rock, which could be overcome only by excessive stress difference.

The conditions favoring rock flowage above summarized seem likely to exist in the deep zone. Within the zone of observation even the strongest rocks have locally suffered rock flowage, and hence have locally, even at that shallow

[1] Adams, Frank D., An experimental contribution to the question of the depth of the zone of flow in the earth's crust: Jour. Geol., vol. 20, 1912, pp. 97–118.

[2] Pfaff, F., Der Mechanismus der Gebirgsbildung, pp. 16–19.

depth, been under containing pressures sufficiently in excess of their crushing strength to permit flowage. With greatly increased pressures at greater depths it is logical to argue that conditions for flowage would be improved. Rock flowage typically involves extensive recrystallization, of a type known from metamorphic studies to mean reduction of volume. Fracturing, on the other hand, increases volume. It is easy for this reason to imagine conditions in the deep zones as favoring rock flowage rather than rock fracture.

Earth temperatures increase with depth. Increase in temperature aids and accelerates rock flowage.

Notwithstanding these and other considerations, any conclusions as to the existence of a deep zone in which all rocks flow when deformed is hypothesis, not proved fact, and perhaps will always remain so. The environmental conditions are not accurately known; and even if each of the factors were measured, their conjoint effect would be speculative. Variations in the time factor alone may determine whether a rock flows or fractures. Rock flowage which has occurred in rocks now accessible to our observation fails to show an increase with depth with sufficient clearness and definiteness to warrant confident downward projection. The increase of rigidity, under the high pressures existing in the deep zone, introduces another doubtful factor. While it is known that, within the limits of laboratory experiment, the rigidity of rocks increases very rapidly with increased containing pressures, the range of pressure capable of management in the laboratory is so small that it does not give satisfactory quantitative demonstration of what the increase in rigidity really amounts to under the high pressures operating deep within the earth. Under these conditions we are warranted in expecting great rigidity, but how great, we do not know. It may be sufficient to resist deforming stresses of almost inconceivable magnitude. Places may exist where the containing pressures make the rocks so rigid that the available deforming pressures

may not be sufficient to cause any movement. Quoting from Adams and Bancroft,[1]

"The experiments seem to indicate that with a containing pressure of about 10,000 atmospheres, which would be equivalent to a depth of about twenty-two miles below the surface, it would be impossible to make the marble flow, except under a pressure which would be simply colossal."

It is this factor of rigidity which makes it so hazardous to calculate the depth within the earth at which rocks might flow on the basis of the narrow experimental data, even if stress differences in the earth were accurately known. The wide range of inferences which have been based on the experimental data by different investigators is eloquent testimony to this effect.

Moreover, the problem is complicated by the fact that experiments with rock flowage thus far have been confined mainly to one kind of stress application, namely, to non-rotational deformation. The results which would be obtained under conditions allowing shearing or rotational movements, for which there is evidence in deformed rocks, have not yet been fully measured. Experiments along this line, now under way at the University of Wisconsin, indicate that under shearing stresses a rock may be made to flow under much less pressure than was used by Adams.

Finally, it is to be remembered that the problem of rock flowage is not merely one of control of pressures, but includes consideration of the time element, the nature of the rocks, the presence of various mineralized solutions favoring recrystallization, and temperature. No one of these factors has been adequately measured in the laboratory, and we have today only a vague idea of their relative importance in producing their conjoint effect as represented in rock flowage. The best information yet available is drawn from a study of the rocks

[1] Adams, Frank D., and Bancroft, J. Austen, On the amount of internal friction developed in rocks during deformation and on the relative plasticity of different types of rocks: Jour. Geol., vol. 25, 1917, p. 635.

which have undergone rock flowage, the metamorphic changes involved, and the environmental conditions which have favored these changes.

Our question, then, as to the extent to which deep movements are accomplished by rock flowage can not be simply and definitely answered in the present state of knowledge. The preponderance of environmental evidence seems to indicate that rock flowage is the distinctive kind of movement, but so many qualifications, definitions, and assumptions enter into this conclusion that our present inclination is to keep firmly in mind the complex facts of deformation in our zone of observation as a possible key to the interpretation of unseen movements. This attitude will require us to pay more attention than heretofore to the possibilities of heterogeneous structural behavior at great depths. Particularly should we keep in mind the fact that the kind of rock flowage accomplished experimentally produces structures which in the earth have sometimes been called fracture, or combined fracture and flowage. We may assume a downward extension of combined fracture and flowage, as observed in the field, and still meet the conditions of flow implied by experiment.

Manner in which stresses are transmitted in rock flowage. The existence of oriented cleavage and shear planes in schistose rocks may be taken as evidence of orientation of stresses in their development. It has also been held to prove the absence of hydrostatic pressures, but this conclusion seems doubtful from certain points of view. Cleavage and schistosity are evidences of movement in definite directions and therefore evidence of oriented stresses, but a very slight stress difference may be sufficient to cause movement in a mass in which the stresses are dominantly hydrostatic. Water will flow in currents toward an opening under its own weight. A combination of differently colored immiscible liquids will become streaked in layers in the direction of movement. These layers are evidence of orientation of stresses, but not of the absence of hydrostatic stresses. In experimental reproduction

of rock flowage, the mass exerts sufficient pressure on the containing walls to stretch them. The resistance to stretching approximately measures the hydrostatic pressures in the mass, and yet there are additional oriented stresses, measured by the difference between the resistance of the walls and the pressure applied by the piston. In any slow-flowing viscous mass there are hydrostatic pressures. If the exterior pressure is applied slowly enough it will be transmitted hydrostatically; if too fast, the viscous materials cannot yield rapidly enough, and some of the stress is transmitted through the mass in the direction of application, while some of it is transmitted hydrostatically. So far as we understand the conditions of rock flowage, as represented by schistose rocks, much the same combinations of stress conditions may obtain.

If there are hydrostatic, as well as oriented pressures, during rock flowage, the question arises how the existence of hydrostatic pressures can be correlated with the known increase in rigidity of the rock mass as indicated by experiment. When a rock is contained, as in the experiments, the stress difference required to make it flow is several times that necessary to crush the same rock with free sides. The rock is said to take on increased rigidity to this extent. At the same time the hydrostatic pressure, absent before, has become sufficient to cause the bulging of the walls. Thus the apparent anomaly that rigidity and hydrostatic stresses are both increased under these conditions. What we mean by increased rigidity is that the mass not only takes on more internal frictional resistance or viscosity, but more resistance because of its inability to move restraining walls. The increase of frictional resistance alone may not be enough to withstand the new environmental stresses, and so far as this quality goes the rock may be actually weaker when considered *relatively* to environmental forces; but when the support given by the environment is also taken into account, the rock is more competent or rigid.

We conclude that rock flowage implies oriented stresses, and probably also hydrostatic stresses, that it implies weakness

relative to environmental forces, but great rigidity in combina tion with them.

It may be noted that rigidity in this sense does not o necessity bear any relation to the ordinary crushing strengtl of the rock. Quartzite or granite, so far as we know, may hav no greater rigidity than marble or slate. Adams' experi ments show diabase and marble in a composite specimen be having similarly. In fact, marble actually penetrated th harder diabase. Likewise, gypsum penetrates steel. Whil there are probably differences in the internal friction or vis cosity of different rocks under these conditions, the results ar nevertheless homogeneous in approximating rock flowage — in contrast to the heterogeneous results under lesser containin pressures, where competency and strength of rocks play a part

Conclusion. Within the zone accessible to observatio movements of rock masses are accomplished by fracture an flowage. These processes may be distinct and separate, or s interrelated as to make definition difficult. The zones o movement are many, their positions and attitudes diverse. I general they indicate shearing or grinding movements betwee rock masses, accomplished by both fracture and flowage, an caused by stresses inclined to the zones of movement. Thi conception is taken to afford the best initial basis for th interpretation and correlation of observed rock structure There is no certain evidence of increase or decrease of move ment toward the bottom of this zone. Beyond a shallo surface zone, there is no certain evidence of increase of roc flowage and decrease of rock fracture with depth. There is n certain evidence that rock flowage means greater weaknes than rock fracture. There is no certain evidence in roc flowage that pressures are dominantly hydrostatic or domi nantly those of competent solid bodies.

Movements are known to occur in the zone below our rang of observation, but their nature and distribution are the sub jects of varied hypotheses based on a few known condition Much of the sharper diastrophism seems to be confined to

thin surficial zone. Deeper movements, of a more massive type, periodic, and possibly slower, seem to be implied by the relative movement of great earth segments as represented by continents and ocean basins. Their depth is unknown. Most of the current hypotheses agree in assuming a single mobile zone in which rocks move dominantly by rock flowage. The basic requirements of reasonable hypothesis, however, may be equally well met by a conception of movement much like that of the zone of observation. This does not require or postulate the existence of any single mobile zone, or zone of slipping, or zone of flowage, or of an asthenosphere. It supposes movement irregularly distributed in many zones, with any inclinations, and accomplished by both fracture and flowage as far below the surface as movement extends — always remembering that some of the structures geologically described as fractures, may be expressions of mass movement of the kind defined as flow in experimental results.

Conditions of temperature and pressure and vulcanism become more intense with depth, but it remains to be shown that their conjoint action results in a uniform environment, and even if it does, that this condition is not upset by what might be called a heterogeneity of the time factor as represented by differing rates of deformation. If homogeneous environmental and time conditions are assumed, it is yet to be shown that these are sufficient to overcome the heterogeneity of the physical properties of the rocks, and to cause homogeneous behavior through any considerable zone. It is not even certain that they may not fix and accentuate the heterogeneous properties of rocks. Certainly in the zone of observation there is comparatively slight evidence of their efficacy in causing more uniform deformation with depth.

In short, as between alternative conceptions as to the conditions in the deep zone, the burden of producing affirmative evidence would seem to rest heavily on any conception involving radical departure from the known irregular distribution and manner of movement within our zone of observation.

CHAPTER XV

CAUSES OF EARTH FAILURE[1]

Cooling of the Earth by Heat Transfer Toward the Surface

The cooling and shrinking of the nucleus faster than the shell causes the shell to collapse, setting up strong tangential thrusts. Notwithstanding this failure as a whole, it is conceived that the rocks are sufficiently rigid to transmit thrusts for long distances, and may be able to form and support not only mountain ranges, but geanticlines and geosynclines, plateaus, continents, oceanic basins, and other large units of structure. This is the old, and present popular conception of earth deformation.

There are three principal hypotheses as to distribution of heat curves in cooling and the resulting deformation.[2]

(1) The earth was once gaseous; then passed into a liquid, and later became a solid. Convection brought the temperature of the whole mass to a uniform point somewhere near the temperature of solidification. A solid crust was formed and solidification proceeded from the surface downward. The crust thereafter did not contract but the part immediately below did. The crust thus became too large for the shrinking interior and collapsed under gravity, developing lateral thrust, which is transmitted in competent or rigid rocks. If the cooling material below is contracting while that above no longer contracts, but is in the state of gravitational compression and tangential thrust, it follows that between the two zones is a

[1] See Chamberlin, T. C., Diastrophism and the formative processes: Jour. Geol., vol. 21, 1913, vol. 22, 1914, vol. 26, 1918, vol. 28, 1920, and vol. 29, 1921.

[2] See Chamberlin, T. C., and Salisbury, R. D., Geology, vol. 1, 1904, p. 534 *et seq.*

level of no stress, where there is neither stretching nor compression. As the cooling proceeds this level migrates downward. With varying assumptions as to initial temperature, speed of cooling, and the period through which cooling has occurred, the level of no stress has been estimated to be at present at various depths down to eight or ten miles below the surface. The shallowness of the zone through which thrust occurs on this hypothesis is taken to accord with the observed facts of the surficial wrinkling of the earth's crust.

(2) Another hypothesis of heat distribution in a cooling gaseo-molten earth takes into account the fact that pressure raises the melting point of rocks; that, if the pressure is high enough, the rock may become solid even though far above its normal melting point at the surface. Little is known of the melting points under such immense pressures as must exist deep within the earth; but it is assumed by extrapolation from the limited range of experimental data that solidification may have begun first near the center of the earth, and thence extended outward. The shrinkage, therefore, instead of being confined to a subcrustal shell, extends through the deep interior.

(3) The earth was formed by the accretion of planetesimals and heat was developed from the center outward by the increase of pressure during its growth. The outward movement of this heat caused the temperature to fall in the lower portions and rise in the outer ones, causing the first to shrink and the second to expand. Curves drawn on certain assumptions as to distribution of density, pressure, and temperature indicate that the outer 800 miles of the earth, or about one-half of its volume, is now under compression or thrust, and is being adjusted by gravity to the shrinking core within.

It will be noted that the first of these three hypotheses of heat distribution concentrates deformation within a very shallow zone. The other two distribute the deformation through deep zones. On the whole the evidence derived from deformation at the surface favors the idea of shallow deformation, though it does not exclude deeper deformation. On the

other hand, the known rigidity of the earth, together with the high pressures and densities within the earth, on the whole, favor the second and third hypotheses.

Changing Rate of Rotation of the Earth

The earth inherited a certain unknown measure of rotation from the sun, and this rate of rotation has been changed by the contraction of the mass, by the infall of planetesimals, and perhaps by other causes. With increasing speed of rotation centrifugal forces tend to make an oblate spheroid. With decreasing speed, the change is toward a sphere. The stresses thus produced are regarded by Chamberlin [1] as the most powerful agencies of deformation to which the earth is subject, and he attempts to show that the segmentation which would naturally result may be identified with the salient features of the earth's surface.

Volume Changes Due to Metamorphism

Earth materials are constantly undergoing cycles of metamorphic and mineralogic change, resulting in considerable changes of volume. The stresses thus set up are known to be the causes of some local deformation. Whether the cumulative effect is sufficient to cause major earth movements is not clear. To a certain extent the metamorphic changes may be regarded as due to chemical and molecular affinities, and therefore as causes of changes in temperature and pressure. It is known also that to some extent they are the results of temperature and pressure changes arising from other causes. When metamorphic changes occur, however, they cause definite physical reactions against the environmental forces. Under suitable conditions of temperature and pressure, earth materials may organize themselves into crystal forms, but this organization is able in its growth to exert pressure and

[1] Chamberlin, T. C., The fundamental segmentation of the earth: Science, vol. 40, 1914, pp. 774–775.

temperature effects on the outside environment and to maintain itself uniformly against an astonishing variety of strenuous environmental conditions, — just as organic life, while a product of environment, also has an independent capacity to react physically on the environment. Looked at broadly, metamorphic changes must be regarded as actual independent causes of deformation, which are known to cause some deformation, and which in their cumulative effect may be one of the great agencies of deformation, but which, in turn, are controlled to a large extent by forces arising from other causes. If these other forces be given the primary place, metamorphic changes may be regarded as one of the agencies through which they find local expression.

Vulcanism

What has been said about metamorphism might be paraphrased to apply to vulcanism. The liquefaction of rocks and the migration of magmas toward the outer surface of the earth are probably the result of changes in environmental conditions of temperature and pressure. The disturbance in equilibrium of these conditions seems to produce both deformation and vulcanism. Very often deformation seems to be a prerequisite to vulcanism; in other cases, it goes on simultaneously; in still others, deformation follows and seems to be caused by vulcanism. The evidence from great batholithic intrusions, as well as other forms of intrusive and volcanic outbursts, is that great pressure is sometimes exerted on the surrounding rocks, even though the origin and movements of the batholithic magma may be in turn the result of changes in equilibrium between temperature and pressure below the surface. To this extent, then, vulcanism must be set out as an independent cause of deformation. If one prefers, it may be regarded as the agency through which the great earth forces find their local expression; but within our zone of observation we see the agency only as a cause. Where erosion has cut

deep, as in the great pre-Cambrian areas, a large part of the deformation may be traced directly to the influence of granitic intrusives. This suggests the possibility, in deformed terranes which are apparently free from igneous rocks, that deeper erosion may disclose disturbing plutonic masses.

Igneous materials cause deformation not only by being driven *en masse* from below, but by volume decrease *in situ* due to the change from a liquid to a crystalline state, and by sudden release of volatile constituents, principally water, which rapidly and sometimes catastrophically increase the pressure (see pp. 249–251).

Continental Creep

The known weakness of rocks under surface conditions, and the evidence of outward movement of earth masses near the margins of continents, have led to the hypothesis that under long-continued pull of gravity continents will slowly creep, spread, and flatten with a glacier-like movement, accomplishing somewhat the same redistribution of surface loads as erosion. Creep may occur by fracture or flow, or some combination of the two. So far as it is accomplished by flow, it is aided by metamorphic processes.

Again, this agency of deformation has complex relationship to other agencies, but so far as its immediate expression is concerned it must be set out as a possible separate agency, appearing as a direct cause in our zone of observation. Whatever its theoretical possibility, it has not yet been established by inductive evidence from observed deformation.

Deformation Due to Isostatic Adjustments

The earth's surface shows some measure of isostatic equilibrium (see pp. 296–298), notwithstanding the fact that erosion and deposition are continuously unloading and loading different areas, thus tending to disturb such equilibrium. It is commonly assumed to follow that the loading and unloading

must be compensated by deep lateral transfers of mass. It is assumed also that the present condition is not a temporary, accidental, or exceptional one, and that adjustments toward isostasy have been more or less continuous in the past.

More specifically it is postulated that unloading by erosion causes areas to rise by elastic expansion of volume, and that deposition of sediments causes an excess of mass in depressed areas, with consequent sinking, and a crowding out of the material somewhere below the lower and denser to the higher and lighter regions, — a movement which would be expressed

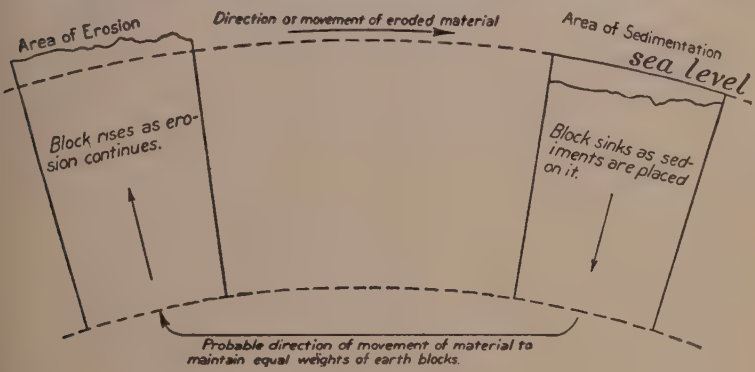

FIG. 103. Illustrating the cycle of movement postulated by the theory of isostasy. The dotted line represents an assumed surface of compensation about 60 miles below the surface. The blocks above the level of compensation are supposed to move vertically as units. The material below the level of compensation is supposed to act as though it were plastic to long-continued stresses. After Bowie.

as a horizontal undertow or underdrag. It has even been supposed that the advance and recession of continental ice sheets might change the load sufficiently to start crustal movements.

There are supposed to be modifying and contributing factors, such as changes of temperature resulting in volume changes, elastic changes in volume under varying load, and metamorphic changes of volume. These factors to some extent oppose each other, and their conjoint effect is not known.

Decrease of pressure by unloading is accompanied by elastic expansion, and increase of pressure by loading is accompanied by compression. It is suggested that the mechanical movement of deep-seated flow will itself increase temperature and that there is further rise of temperature in areas of recent deposition, because heat coming up from below is prevented by the new blanket of deposits from rising from the surface so freely as before, or because the bottom of the column sinks into a hotter region. Conversely, beneath the areas of recent erosion the temperature of material at a given depth will decrease. As increase of temperature increases volume, it is apparent that the temperature changes assumed will tend to some extent to lessen the elastic volume changes produced by the loading and unloading. Relief of pressure by erosion causes metamorphic changes which are accompanied by increase of volume and reduction of density. The reverse series of metamorphic changes goes on with increasing load. The volume changes resulting from metamorphism are on the whole in an opposite direction from those produced by variations in temperature, and in the same direction as the elastic changes due to variations in load.

The movement postulated by the theory of isostasy is obviously not to be regarded as a definite one, upon which all isostasists agree. In fact, almost no two agree in all particulars. In the various views the undertow is thought to take place above the depth of compensation, below the depth of compensation, at both depths, in the same direction above and below, or in different directions above and below. The rôles of temperature, metamorphic, and elastic volume changes are variously emphasized. Likewise, there is diversity of opinion as to how far any disturbance of isostatic equilibrium can be regarded as a primary and independent cause of deformation, and how far it is subsidiary to other great causes, such as the shrinkage of the globe.

The movement cycle postulated is attractive in its simplicity, and for the time being at least is regarded by most

geodesists and a good many geologists as one of the principal agencies of deformation. Another point of view is that deformation is principally due to more fundamental, underlying causes; that disturbance of isostasy by erosion and deposition is only a subordinate and contributing cause of deformation; and that the particular scheme of movement supposed to result from it may not be a reality.

Critical consideration of this subject falls naturally into two parts: first, the geologic difficulties involved in assuming that disturbances of isostasy by erosion and deposition are a primary cause of rock movement; and second, the manner in which the condition of isostasy may be related to other major causes of movement. In both cases there is acceptance of the fact that the surface of the earth shows some measure of isostatic balance, though with reservations which have been noted in a preceding chapter. It is also assumed that this condition may have persisted in the geologic past, though for this there is no direct proof.

Disturbance of isostasy by erosion and deposition as a cause of rock movement. The cycle of movements supposed to result from disturbance of isostatic equilibrium presents several difficulties to the geologist. Most of these have been recognized by geodesists, but in some quarters there is perhaps a tendency to minimize them.

1. The theory postulates a marked difference in behavior in rocks above and below a hypothetical depth of compensation. Below they are weak enough to allow continuous undertow by rock flowage. Above they are asssumed to have sufficient rigidity to maintain columns of different density and to support inequalities in topography. The movement between them in the upper zone is supposed to be dominantly vertical; their resistance to vertical load is believed to be less than to horizontal compression.[1] If the columns of different density were not sufficiently rigid to prevent lateral move-

[1] Wood, H. O., Some considerations touching on isostasy: Bull. Geol Soc. Am., vol. 33, 1922, pp. 303–316.

ment, there is no apparent reason why above the depth of compensation there should not be movements from columns of low density to columns of high density, — the low density of the taller column being more than offset by its increased height. If, to illustrate, a level surface of reference, a mile below sea bottom, be considered, it is obvious that the weight on this surface would be less under the sea than under an adjacent mountain mass, — where the rocks, though lighter, stand several miles higher. The movement called for is clearly one from beneath the mountain to beneath the sea; and it is only by assuming a high degree of lateral rigidity above the depth of compensation that the absence of such a movement can be explained.

2. As a corollary to the emphasis on dominantly vertical movements between the columns, it is necessary to minimize tangential shortening or thrust as a cause of uplift, because such thrust would cause an excess of load in the uplifted area.

The general isostatic scheme of rock movement cannot in the nature of the case be proved by observational evidence, nor can it be denied on this basis. It is an artificial and systematic scheme which contrasts with the observed heterogeneous behavior of rocks in the zone of observation. As indicated elsewhere (Chapter XIV), the writer is inclined to base his conceptions of deep rock movement on the kinds of movements known in the zone of observation. These are the only movements we actually know anything about, and from our viewpoint the burden of proof falls on any attempt to present a scheme of movement which departs radically from the observed facts near the surface. In the zone of observation movements are known to be vertical, horizontal, and inclined, and it seems to be rather a violent assumption that immediately below this zone movements should be largely of a vertical nature, and still further down dominantly horizontal. Thrust or tangential shortening is known to be an important mode of deformation in the zone of observation, even though it is well recognized that vertical uplift, without

shortening, is also an important process. While rocks at the surface may be too weak to carry great thrusts of the kind necessary to produce mountains, where rocks are under load their greatly increased rigidity seems to make them competent for the work. At any rate the evidences of thrust exist. Calculation of the depths of masses affected by mountain-making movements seems to indicate beyond reasonable doubt that tangential shortening extends far below our zone of observation (see pp. 241–249). The structural geologist is naturally skeptical of any hypothetical scheme of movement which minimizes its importance.

Moreover, it seems hardly logical to assume that lateral transfers of load by erosion at the surface are automatically compensated by volume changes in the rocks below, and to imply that lateral transfers of mass by thrust would not be likewise compensated.

It is true that the question as to the part played by lateral thrust or tangential shortening in the major deformation of the earth has been more or less confused by the assumption that the shortening evident in mountain chains was contemporaneous with the uplift and the cause of uplift. It is true also that in some mountains uplift has followed or preceded deformation caused by tangential shortening. This shows nothing more than that crustal shortening is only one of the phases of earth deformation; it does not disprove its reality.

3. The existence of a depth of compensation at 60 miles, or at any uniform depth, is concededly speculative, and yet this assumption is basic to a belief in the existence of a single zone of rock flow at some assumed depth. If the vertical distribution of density is irregular, there is no apparent reason why rock flowage should not also have irregularity of vertical distribution. If density really changes gradually to the center of the earth, then the depth of compensation might be located at the center — in which case the zone of rock flow and compensating underdrag reduces to a point.

4. The hypothesis of isostasy calls for a continuous, slow

movement under stresses of small magnitude in order to explain the present delicate adjustment in the face of continuous erosion and deposition. So far as the strength of the earth's crust is known, it seems to be more than sufficient to withstand any stresses calculated from depths of unloading and loading. Quoting from Willis: [1]

> "According to Adams's experiments, the rocks at a depth of 5.8 miles below the surface have so gained in strength, because of confining pressures, that they are competent to support a column of rock 23.3 miles high; that is, they are four times as strong as the load which rests on them, and, even though the weight of a mountain range two or three miles high were added or subtracted, there would be no movement in the rigid foundation, except an elastic one."

As noted elsewhere (p. 291), these calculations have not been based on experimental determinations of the strength of rocks under shearing stresses, and the question of the strength of the crust cannot be regarded as settled.

Admitting the rigidity of the earth under suddenly applied stresses, the geodesist then argues that it may be weak and plastic under long-continued stresses. This meets another difficulty, for there is overwhelming geologic evidence that earth movements have not been slow and continuous, but have been periodic in their nature. There have been long periods of quiescence, during which stresses presumably have accumulated, followed by periods of deformation during which stresses have found relief.

5. In general it may be said also that the theory of isostasy fails to provide adequate causes for reversals of movement, and for lateral shift of movement zones, both of which are established geologic facts. Changes in temperature, with consequent volume changes, are supposed to be at least one of the causes for reversals of the direction of movement. Heavy sedimentation is supposed to cause sinking and a consequent slow rise of temperature in the lower part of the

[1] Willis, Bailey, Rôle of isostatic stress: Bull. Geol. Soc. Am., vol. 33, 1922, pp. 372–373.

column, which will cause expansion and the ultimate uplift of the column. Similarly when an area of erosion rises, the temperature of the lower part of the column is supposed to fall, resulting in reduction in volume. After erosion has stopped, due to base-leveling, the cooling and reduction in volume are supposed to continue, on the assumption that the earth's isogeotherms were temporarily raised during the uplift and slowly fall to a normal position after erosion has completed its work. This situation would tend to produce a depression where formerly there had been an uplift. It is recognized that change in volume due to this cause is not quantitatively sufficient, judging from experimental results, to cause the known elevations and depressions, and that some other causes may be operative, but these causes are not specified by the geodesist. This general argument also ignores a vital fact, — that the actually observed metamorphic changes of volume under loading and unloading are in an opposite direction from those due to the supposed temperature changes, and that these changes are of greater magnitude. Thus the loading of delta formations changes a mud to a shale or slate, with elimination of pore space, water, oxygen, carbon dioxide, and recombination of the remaining constituents into relatively heavier silicates, — resulting in a volume reduction which, judging from the evidence available, is of considerably greater importance than any possible volume increase due to rising temperature.

6. The volume changes assumed in the theory of isostasy are usually ascribed to changes of temperature, elastic expansion under changing load, metamorphic changes, and so on, leaving the impression that these changes are common to and of equal magnitude in all rocks, and that these factors would accomplish the known distribution of densities even though the material upon which they act were homogeneous. Comparatively little stress has been laid upon the fact, established by Washington,[1] that differing densities are to be correlated

[1] Washington, H. S., Isostasy and rock density: Bull. Geol. Soc. Am., vol. 33, 1922, pp. 375–410.

with different kinds of igneous rocks, which inevitably make up a large proportion of any zone of movement involved in isostatic readjustments. The high areas of low density are made up of rocks differing in kind from the low areas of high density. The actually existing distribution of densities has been determined largely by the migration of magmas, or by whatever conditions have controlled their distribution (see pp. 299–300). If, therefore, the loading and unloading by erosion and deposition have set up stresses sufficient to cause movements in the underlying rocks, and these movements are in the direction of restoring isostatic equilibrium, it follows that the migration of magmas is likewise controlled by this agency. Such a possibility is hard to disprove; but in geologic thought the origin and movement of magmas have been so largely associated with other great causes of earth deformation, rather than with the mechanism of isostasy, that the old conception is not likely to be easily abandoned.

7. In general, if disturbance of isostatic equilibrium by erosion and deposition be regarded as a primary cause of deformation, no satisfactory way has been found to correlate it with other known causes such as crustal shortening or igneous intrusion. The mechanism postulated seems to require the field for itself, free from complicating movements due to other causes, some of which would have an opposite effect.

8. Finally, the discussions of isostasy place much reliance upon a concealed assumption, namely, that the shifting of load by erosion and deposition is the primary *cause* of subsurface movements compensating for this shift. It is concluded that the earth cannot be rigid because it is sinking under load. There has been nothing like adequate consideration of the possibility that the surface shifting of materials and sinking of geosynclines may be the *result* of movements from other causes, that the shifting may be tending to restore rather than disturb isostasy, and that the distribution of densities, upon which the theory of isostasy is based, may be

the result of quite a different scheme of movement. These possibilities are discussed in the following sections.

Manner in which the condition of isostasy may be related to other causes of movement. (1) Uplift may be caused by lateral shortening, or by vertical movements due to plutonic intrusion or other causes. To make this discussion as simple and definite as possible, let us start with the known fact of tangential shortening, however caused, — whether by cooling and shrinkage of the earth, by changing rate of rotation, or in any other way noted in this chapter. Let us assume for purposes of discussion that the earth's surface was initially level and that there was uniform distribution of densities. Thrust or folding would then cause excess of mass in uplifted portions. The ensuing erosion and deposition would tend to restore isostatic equilibrium, rather than to destroy it as postulated by the theory of isostasy. While there may be more or less of a balance between speed of uplift and speed of erosion, there is a lag measured by the elevations on the earth's surface at any time. These may be supposed to represent an excess of mass, which erosion has failed to reduce.

But there are several factors other than erosion and deposition tending to equalize densities. It is known that uplift by compression is often associated with great batholithic intrusions of granitic rocks, which are light and tend to lower the density of the uplifted mass.[1] Many of the cores and higher peaks of the Himalayas, Alps, and Rocky Mountains are dominantly granitic. The roots of old mountain ranges disclosed by deep pre-Cambrian erosion are also dominantly granitic, and furnish contributory evidence of the common association of granitic intrusion and uplift. For the purposes of the present argument it does not matter whether the intrusion be regarded as cause or effect, or both, of tangential shortening (see pp. 249–253).

The elevated mass undergoes elastic expansion; openings develop by fracture; the mass undergoes katamorphism, with

[1] Cf. Washington, H. S., *loc. cit.*

accompanying expansion of volume. Also, it may be supposed that some of the energy expended in crustal shortening is translated into heat, which causes expansion of the rocks, and therefore the lowering of their density. It may be supposed that this heat would be localized in regions of most intense deformation, which in turn correspond to the mountainous regions of uplift. This factor, however, might be offset by uplift of the mass into cooler regions. The weight of the elevated mass is not lessened by these processes, but the decrease of density may affect the readings at nearby stations, and cause apparent compensation.

So far as there are original differences in density inherited from the early stages of earth development, the problem is complicated. In this case tangential shortening and uplift might either restore or destroy isostatic equilibrium. So far as uplift was concentrated in the originally light areas it would tend to restore isostatic equilibrium. It is possible that the originally light areas were also the weak ones, which would tend to localize movements in these areas, but as indicated in Chapter XIII lightness and weakness are not necessarily synonymous.

In general, uplift by tangential shortening may be supposed to disturb any existing isostatic equilibrium, just as erosion and deposition would disturb it, but the movement inevitably starts a series of agencies all tending, actually or apparently, to restore it — erosion and deposition, granitic intrusion, expansion of volume due to metamorphic changes and to heat generated by mechanical movements. We cannot say, on the basis of full quantitative information, that these changes are sufficient to compensate for the excess of mass caused by compressional uplift, but the information available constitutes a formidable obstacle in the way of asserting that they do not. It is to be emphasized that conclusions as to the completeness of existing isostatic balance also include certain assumptions which prevent final quantitative results. There is no question about the reality of some degree of isostatic balance; neither

is there question about the existence of factors tending to restore isostasy in elevations caused by compression; the exact quantitative value in either case is yet unknown.

From this point of view the condition of isostasy may be regarded as the result of the conjoint action of several more or less mutually balancing factors, of which erosion and deposition are merely one.

This conception of the maintenance of isostatic adjustment involves no hypothesis as to the existence of a single zone of undertow or flow deep below the surface, or maintenance of rigid columns moving vertically above this zone. The movements may take place at all depths and by any combination of fracture and flow. They may be tangential, inclined, and vertical. They involve mass transfers at any depth into the zones of compressional shortening. The movements stop when tangential shortening stops. They do not set up a cycle which will be continued by loading and unloading, or by changes in density.

The localization of uplift under compressional shortening would be largely a structural one due to initial dip and to weakness in rocks (see pp. 193–194). The migration of uplift zones, the reversals and renewals of uplifts, so difficult to explain if erosion and deposition are the primary causes of movement, are natural and inevitable expressions of crustal shortening.

The fact that mountainous uplifts are often, though not always, localized over great geosynclinals of heavy deposition may be explained by the initial dips in these synclines, by the steep slopes of contact of sediments and basement, by the contrast in strength between the soft sediments and the adjacent land areas, and so on. The sinking of the geosyncline need not necessarily be regarded as the result of loading. It might be started by tangential shortening or other causes. Reid [1] has called attention to the fact that great

[1] Reid, Harry Fielding, Isostasy and earth movements: Bull. Geol. Soc. Am., vol. 33, 1922, p. 317.

depressions have taken place without the addition of sediments, as for instance the great deeps of the ocean, and as a particular example the Tonga Deep, which are so situated that they never could have received any large amounts of sediments. Also there are great geosynclinal deposits of ancient age which have not later been elevated into mountains.

Erosion and deposition certainly change the distribution of competency in rocks by destroying competent structure, by changing the load, and by piling up structurally incompetent masses of unconsolidated material, and in this way may cause an outlet for stresses existing in the rocks below. It is not easy to see just what the effect on the distribution of deformation might be. If the underlying stresses were working toward the continued uplift of an elevated mass, erosion would tend to help them by taking away load, but this help might not be sufficient to keep the movement going in the face of the thickening and strengthening of the mass resulting from tangential shortening. The forces of shortening in time would find it less easy to lift the elevated mass farther than to start a new attack on an adjacent area which had not been thickened, and which furthermore had been weakened by the accumulation of soft materials, as in a geosyncline.

(2) Instead of starting with crustal shortening as a primary cause of deformation, let us assume that development or migration of magmas may be the disturbing element, with or without folding. This is usually related in some way to earth deformation, as cause or effect, or both. But for whatever reason magmas have formed, it seems to be a fact (already noted, pp. 299–300) that they have come to rest in such positions that the lighter ones constitute the elevations on the earth's surface and the heavier ones the depressions. The distribution of igneous masses, both vertically and horizontally, presents a very remarkable correlation with the distribution of density inferred from gravity determinations. Their density has apparently determined their vertical positions.

[1] Gilbert, G. K., Geology of the Henry Mountains: 2nd ed., Washington, 1880.

Gilbert [1] long ago established a relationship between the densities of the different Henry Mountain laccolites and their vertical positions in the section. There are, of course, important modifying structural features, such as the competence of the overlying rocks to resist the upward course of the magmas.

Under the general conditions here postulated erosion and deposition would tend to destroy isostatic equilibrium, and it would be necessary to assume that the maintenance of equilibrium requires either the influx of new magmatic material from below or the pushing up of the cover under the urge of still molten masses below struggling for gravitational adjustment. It would be necessary to assume, furthermore, that these opposing forces be more or less in balance, though not exactly so, for there are considerable variations from complete isostasy to be accounted for.

As under any other hypothesis, there would be other contributory factors such as elastic expansion under unloading, expansion under katamorphism, reduction in volume through cooling, and so on. Cooling in this case might be supposed to cause a considerable reduction in volume.

Under this hypothesis the rôle of erosion might be that of directing the movement of the magmas, in so far as covers which prevent their upward migration are removed or weakened by erosion. Also accumulation of unconsolidated sediments might furnish conditions for magmatic invasion. Erosion and deposition might supply the proper conditions of relief for the expression of forces acting from below and originating from some other cause. That in themselves they could supply great enough forces to cause liquefaction and movements would seem highly doubtful.

The cessation of upward movement in one area and its transfer to another area which was formerly one of heavy deposition is due to causes about which we can only speculate; under this hypothesis it might be assumed that the cooling and consequent crystallization of an upward-moving magma in time yields a cover too competent to move, and that the

magma may then find structurally incompetent zones in adjacent areas of soft sediments to break into and lift.

Conclusion as to Ultimate Causes of Earth Failure

The principal causes of deformation, so far as they are yet surmised, are summarized on preceding pages. While all of these are plausible causes of deformation, the actual effect of no one of them has been measured, much less their conjoint effects. So far as it is possible to generalize from this vague state of knowledge, it may be said that geologists are at present inclined to give principal place to changing rate of rotation and to the shrinkage of the earth, due to heat transfer from the interior outward, whether they go back to the nebular or planetesimal hypothesis of the origin of the earth; that metamorphism and chemical changes, vulcanism, and forces tending to maintain isostatic equilibrium are regarded as subordinate or contributory causes, or perhaps as special and local expressions of the more basic causes first indicated.

Erosion and deposition, by disturbing isostatic balance, may start a cycle of adjustments, consisting of: depression of loaded vertical columns, acting as units without lateral flow, above a depth of compensation; below this, a lateral flow in a hypothetically weak zone from beneath the loaded column to beneath the higher column; vertical uplift of the higher column. This involves a series of assumptions as to behavior of rocks at different depths which involves difficulties from a geologic standpoint.

The facts of isostasy may also be explained by a combination of other causes, which do not require the movement cycle following from the assumption that erosion and deposition are the controlling factors. The primary place may be taken by crustal shortening, due to gravitational adjustments of the earth dependent upon heat transfer, changing rate of rotation, or other causes; or the uplift may be merely a vertical one, due to plutonic intrusion or other causes. Whatever

the cause, the uplift and excess of mass so caused are in large part automatically compensated by a lowering of the density through elastic expansion, breaking, katamorphic changes, and beating due to mechanical movement, and especially by accompanying migration of igneous masses. Erosion and deposition may sometimes cause localization of deformation due to other causes, but they are not always competent to do so; their effect is only subsidiary; they may not be able to start a chain of movements of their own independently of other causes. The cycle of movement postulated by the theory of isostasy requires freedom from interference by movements due to other causes, and cannot be made to accord with other movements. While in some cases, as Willis [1] says, "isostasy is the rudder, not the motive power, of deformation," in others its rôle may be even less important.

While the writer leans toward the latter explanation of isostasy and the movements connected with it, the student is cautioned against full acceptance of any simple and definite explanation in our present state of knowledge. The problem includes so many unmeasured and perhaps immeasurable factors that no living scientist can claim even an approximately correct perspective; all are groping for the light.

[1] Willis, Bailey, Rôle of isostatic stress: Bull. Geol. Soc. Am., vol. 33, 1922, p. 374.

CHAPTER XVI

FIELD METHODS

The separate treatment of the different aspects of structural geology in preceding chapters has not allowed the presentation of certain general considerations touching field methods. The student with a book knowledge of miscellaneous structural details is likely to feel at a loss in planning his attack on a complex structural field problem. No two problems are exactly alike, and there is no fixed procedure which can be indiscriminately applied. There are, however, certain broad lines of approach more or less common to a great variety of problems, which may be briefly presented in perspective.

1. First is the accurate observation and recording of the facts in such a manner that the record itself will stand, no matter how much the inferences from them are subsequently changed. Inferences should be so clearly separated from facts that anyone reviewing the work will readily understand just how much is actually known and how much is surmised. Much good field observation has been vitiated by failure to observe this rule.

2. Rock structures should be studied and mapped in three dimensions. It is good practice to plat the observations on three mutually perpendicular sections, or to express them on a three-dimensional block diagram, or to use contours to indicate structural forms, especially folds. It is not often that field exposures exhibit all three dimensions; usually the observations relate to only one or two sections, and the effort to picture the facts in the third may lead to discovery of new features or to a better understanding of the structure as a whole. For illustration, field work may lead naturally to the

platting of a fold in horizontal plan and next to platting it on a vertical section normal to its axis. Its platting on a section parallel to the axial plane may yield still further desirable information.

3. The platting should be completed so far as possible while on the ground. The attempt to do this nearly always discloses omissions or doubtful features, which can then be checked over. When the field notes are hasty and sketchy, it is difficult later in the office or camp to put information into proper form. Details are forgotten, the perspective more or less changed, guesses are made as to omissions and doubtful points, and there is more chance of error in transferring the notes to maps.

It is especially desirable that structural details be platted accurately in regard both to scale and direction. The liberal field use of the protractor may bring out relationships not suspected from casual inspection of data recorded only in notes. The underground mapping of mines, especially in a much faulted district, can scarcely be done by other methods.

4. The relations of different structures should be studied. A single deformation may express itself in a great variety of structures, and each of them may throw light on the others and upon the deformation as a unit. The typical relationships described on preceding pages should be looked for. If there are departures from the normal relationships, an attempt to discover the reason often leads to the recognition of new factors in the problem. The outstanding fact of earth deformation is the heterogeneous response of different kinds of rocks to stresses, and the great variety of stresses locally involved in any major movement. The results of deformation are not uniform through large areas. Folds, joints, faults, fracture cleavage, flow cleavage, and larger units of structure may be simultaneously or progressively developed in a single great movement. Inferences drawn from any one of them should be checked by inferences from the others. The displacement shown on a single fault plane or the attitude of a

single fold may be quite misleading as to the nature of the major movement, due to the fact that the major forces may be locally translated into all sorts of apparently contradictory movements. It is only by correlation of evidence from all different structures that a clear picture of the deformation may be secured.

5. The evidence of the time relationships of the different structures are to be noted — whether folding and cleavage are contemporaneous or of different ages, whether joints are related to folding or have formed later. Failure to recognize the difference in behavior of different formations under the same movements has sometimes led to the assumption that they were deformed at different times by different forces.

There has been rather a widespread assumption among geologists that rock deformation has been accomplished in the hard-rock stage, and structural features have been interpreted on this assumption. It is known that many structures date back to a period when the rock was soft, molten, or unconsolidated.

6. Any inference as to the source of ultimate forces and their direction of application should be made with care and with understanding of the great variety of possibilities. The forces may be local or regional, uniform in direction or highly diverse. The directions of shortening and elongation of the rock mass may often be accurately noted, but it is only under very special conditions that it can be determined how the deforming stresses originated, whether the elongation was normal to the pressure or inclined to it, and if inclined, what the direction of pressure was. Geologic literature discloses much confusion due to the failure to discriminate between shortening and elongation on the one hand, and the direction of application of stresses on the other. For instance, shortening by folding is ascribed to a pressure normal to the axial plane, whereas the same result might well have been secured by shearing stresses oblique to the axial plane (see Fig. 77).

7. The procedure above outlined starts with the accurate

recording of details and proceeds inductively to the consideration of large elements of structure, their causes and forces. In some fields, exposures at the surface and in mines are so good that more or less mechanical routine observation may require only a minimum use of structural principles. It is simply a question of careful three-dimensional mapping. But in most fields there is so much hidden that it is practically impossible to do effective field work without constant use of structural principles; in the nature of the case, working hypotheses must be employed, no matter how much one may desire to confine his efforts to direct observation. The facts may be recorded separately from the inferences, but exposures are seldom complete in three dimensions and some projection or inference is necessary to complete the story. Hypothesis sharpens the perception and aids observation. With additional hypotheses to test, more things are seen. No eye can take in at once all of the possible facts of observation, unless backed by omniscience. Most geologists have from time to time been surprised in revisiting localities, where they had supposedly noted all the essential structural facts, to find how widening comprehension of the origin of the structures and general principles of the subject disclosed to their eyes many significant features which they had previously overlooked entirely, or had noted as insignificant details and exceptions.

In using this method there are two extremes to be avoided. At one, the observer is so completely convinced of the truth of a hypothesis that he shuts his eyes to everything not in accord with it. At the other, the observer has no conscious hypothesis, and attempts to set down only what is actually observed. With the best intentions, he usually does not see all the features which should be recorded. Some details are emphasized, others overlooked. As a matter of fact, the so-called "practical" observer has his full complement of working hypotheses, though he does not consciously designate them by this name.

Good observation requires constant shifting between the in-

ductive and the deductive methods, the use of multiple working hypotheses, the training of the mind to the greatest variety of possibilities, the impartial search for evidence, the ready abandonment of hypotheses which do not work. The fear of being regarded theoretical should not be allowed to curb the freest use of the imagination in the search for the truth, but the fruits of imagination are not to be set forth as facts or conclusions without full evidence.

When, after elimination, one hypothesis seems to be the most probable one, this should be checked by following it through to its logical conclusion. It is the "follow through" that counts, as it does in golf or billiards. By this we mean that we must be careful to ascertain the full consequence of any hypothesis adopted. If, for instance, there is reason to suspect, from facts observed on a vertical cross-section, that a certain set of joints is developed by a shear acting in a certain direction, care should be taken to see whether the facts observed on the horizontal section or on another vertical section at right angles to the first check out with this hypothesis. Very often this attempt to "follow through" discloses new factors which have been overlooked, and requires modification or amplification of the preliminary hypothesis.

The question may naturally arise in the mind of a student whether it is not better to stop with a description of facts and not bother about difficult questions of origin. The geologist who confines himself to this limited field is doing a small and limited part of the necessary work, and much of this could be almost equally well done by the engineer or the layman who is a good observer. With an understanding of the general nature of rock deformation and the principles involved, the geologist should be able to go a step farther, gaining a better understanding of the structural situation as a whole and a better discrimination between essential and non-essential facts of observation. In economic work the structural geologist is called in, not solely because he can make an accurate strike and dip measurement — the engineer can do this just as well

— but because of the possibility that he may be able to discern in the given situation some controlling principles or salient features which will eliminate much useless and expensive exploration.

8. There remains to be discussed another feature of structural field work, which is fundamental but difficult to express. This might be called "scale" or "perspective." It is practically impossible to record all of the detail in any district. The beginner may start with commendable zeal to record all he sees, and may soon be lost in a maze of detail. He fails to discriminate essential from non-essential details and to see some of the important larger features. It is highly desirable before systematic mapping is started to take a general look over the field, to see what structures there are, to decide so far as possible which features shall be selected for especial attention, the detail and scale to be used, and to select the methods of attack best adapted to the given case. Blind adherence to a set procedure wastes time. A set of observations useful for one field or formation may be superfluous for another. For instance, strike and dip observations are essential to structural mapping where folds are open and of some magnitude. When this method is carried over into closely contorted areas, where individual folds may be measured in inches, strike and dip observations become practically useless, because of the almost infinite variety of results obtainable. Efforts should then be concentrated on the directly observable pitches and the attitudes of the axial planes of the folds. Students who attempt to map the intricately folded jasper peaks on the Marquette and Vermilion Ranges of Lake Superior soon come to realize that a million strike and dip observations would not tell the story, whereas attention to the axial planes and axial lines soon discloses an interesting single unit of folding which can be effectively used in determining the structure of a region. Minute joints which exist in thousands in certain formations may be only incidental structural detail in one district, and the huge effort necessary

to record them may be superfluous. In another district where such structures are mineralized by ores of copper and precious metal, the utmost detail may be necessary.

Attempt should be made to relate structures to some larger unit of structure. The platting and recording of a series of separate joints should be correlated and integrated, to see if there is any definite system of fracture. The local attitude of cleavage may be a little detail of folding. Small folds may be merely drags on the limbs of larger folds, and these in turn the detail of still larger structures. It is necessary to build up constantly from the smaller to the larger features. The effort to record the detailed facts faithfully often stops short of consideration of larger elements of structure. After working weeks and months on some structural unit, one suddenly wakens to the fact that this is but a part of a larger and still more interesting story, and wonders why he had not thought of it before.

The controlling or salient feature of structure is to be looked for. In one case it may be folds; in another it may be faults or joints; in another cleavage. In each case the dominant structure is likely to be accompanied by other structures which are of subordinate importance. Or the dominant structural feature may be the differential movement of one great section of the lithosphere past another, considered vertically or horizontally. When the detail of individual structures is looked at, the general structure may seem too complex for solution. When the separate structures are regarded as minor and incidental expressions of a single great movement they may take on order and significance.

In some highly deformed terranes, like the lower divisions of the pre-Cambrian, there is such a great variety of rock structure that the attempt to unravel it seems almost hopeless; but even here it is sometimes possible to note that the folds are predominantly of a drag type, indicating differential movement in a certain definite direction,— which can be checked by noting the prevailing differential movement shown by joints,

faults, cleavage, etc. The main feature then to be noted is the inclination and direction of the plane or zone along which the drag has occurred. This then becomes a unit which can be used in getting at some of the larger features of the structural problem. The platting and description of any one of the local detailed structures might be almost meaningless, but the unit drag which they indicate is sufficiently definite and extensive to be of some general significance.

The above suggestions for procedure in structural field work are by no means inclusive; experienced geologists will think of many others. The writer has naturally presented certain ones suggested by his own field experience. The dominant note is the use of multiple hypotheses, with variations in method of attack. Observational routine adapted to one district may be useless in another. A man trained in one field has to recast his methods in another. Sometimes his habits have become so fixed that he is almost useless in another field. The essentials are insight, imagination, elasticity of attack, capacity to view the subject broadly and objectively, and the selection of the right methods for the given problem.

APPENDIX

SUMMARIES OF EXPERIMENTS ON STRUCTURAL GEOLOGY[1]

For the following summary the writer is indebted to Dr. Edward Steidtmann of the University of Wisconsin. The list is not exhaustive, but is supposed to be fairly representative of the better known work along these lines.

A. Joints and Faults

The experiments on the strength of materials which could be considered in connection with the origin of joints and faults are exceedingly numerous Most of the experiments referred to below have been made specifically with reference to the problem of joints and faults.

Lüders[2] observed that both sides of an iron bar where bent showed closely spaced lines crossing each other at nearly 90°, in directions diagonal to the axis of the bar.

Daubrée[3] developed intersecting joints by twisting glass plates. The joints formed were nearly at right angles to each other and roughly 45° to the axis of torsion. The compression of blocks of wood gave rise to sets of joints making angles of about 45° with the axis of compression.

Cadell's[4] experiments deal with the problem of repeated faulting of the northwestern Highlands of Scotland. The faults common to this region are similar in strike and direction

[1] In a paper on demonstration material in geology, Cleland has described a number of bits of simple apparatus and models for use in teaching structural geology. (Cleland, H. F., Demonstration material in geology: Bull. Geol. Soc. Am., vol. 33, 1922, pp. 56–84.)

[2] Lüders, W., Über die Äusserung der Elastizität an stahlartigen Eisenstäben und über eine beim Biegen solcher Stäbe beobachtete Molekular bewegung: Dinglers Polytech. Jour. Bd. CLV, p. 18, 1860.

[3] Daubrée, A., Études Synthétiques de Géologie Expérimentale, 1879.

[4] Cadell, H. M., Thrust faults without folding — Experimental researches on mountain building: Trans. Roy. Soc. of Edinburgh, vol. 35, 1888, p. 337.

of dip, but each in turn is less inclined (proceeding in a direction opposite to the dip).

Cadell used a box with a push block at one end, controlled by a screw. The layers used, stucco, sand, and clay, rested on wood. No load was applied.

An arch usually formed directly back of the push block. After rising to a certain height it was either followed by other arches farther back or by thrust faults on the limb away from the thrust. Brittle strata commonly formed thrust faults without preliminary folding. Lateral, upward, and downward gradations of thrust faults into overturned folds were observed. A common result simulating Highland structure was the piling up of a group of slices formed by a succession of thrust faults dipping in the general direction of thrust, but each in turn somewhat less inclined. Finally the whole pile would ride forward on a low-angle thrust plane.

Crosby [1] considered earthquake waves as important in rupturing rocks already under strain. To test this view, he twisted a glass plate nearly to the breaking point by Daubrée's method. The stress was then somewhat relaxed and the torsion rod placed upright on a table. The glass was easily cracked by vibrations caused by hammer blows on the table. The joints produced were like those in the experiment of Daubrée and were not affected by the directions from which the vibrations came.

Becker [2] repeated Daubrée's experiment of twisting a glass plate to develop intersecting joint systems. On thick plates ruptures follow curved surfaces which outcrop in straight lines on the surface from which they begin. The two sets start on opposite sides of the plate.

Hartmann [3] published experimental data on the fracture of metals resulting from various stress conditions.

[1] Crosby, W. O., The origin of parallel intersecting joints: Am. Geologist, vol. 12, 1893, pp. 368–375.

[2] Becker, G. F., The torsional theory of joints: Trans. Am. Inst. Min. Eng., vol. 24, 1894, p. 130.

[3] Hartmann, L., Distribution des déformations dans les métaux soumis a des efforts, Paris, 1896.

Sheldon [1] developed intersecting sets of compression joints by squeezing paraffin bars strengthened with resin in a vise.

Chamberlin and Miller [2] studied the cause of low-angle faulting experimentally. For compressing layers they used a box of the Willis type which had a push block at each end. The layers used were made of various mixtures of clay, sand and plaster of Paris. A strong layer between weak layers rose to a simple arch which broke at the top. A group of strong beds between weak beds broke along rather steep planes which dipped both toward and away from the push block. Movement on these breaks led to both over- and underthrusting. With layers in which differences in competency were moderate a low-angle thrust, dipping about 20 degrees, developed near the surface and became nearly flat in the weak layers below. Farther down it steepened slightly. With material piled high near the push block, a series of steep faults was obtained near the push block, each in succession forming farther from the push side and at a somewhat lower angle than the one before. Paraffin blocks under rotational strain caused by a horizontal endwise push against their upper portions broke along curved low-angle surfaces.

Quirke [3] made several experiments to show some factors controlling high and low-angle thrust faulting in beds. He compressed sheets in a vise. Rotational stress was caused by placing a little wedge, large end up, between the jaws and the test piece. Cakes of soap broke at the bottom along a low-angle fault which flattened out near the middle and increased to 60° near the surface. Narrow pieces of soap broke on steep surfaces near the edges. A plate of soap wider than long broke on a low-angle plane only. In paraffin blocks which bent but slightly, the fractures were usually steep.

[1] Sheldon, Pearl, Some observations and experiments on joint planes: Jour. Geol., vol. 20, Nos. 1 and 2, 1912.

[2] Chamberlain, R. T., and Miller, W. Z., Low-angle faulting: Jour. Geol. vol. 26, No. 1, 1918, pp. 1–44.

[3] Quirke, T., Concerning the process of thrust faulting: Jour. Geol., vol 28, 1920, pp. 417–438.

Quirke believes that in sheets under rotational strain which bend before breaking the fractures are steep at the top because of tension, and flatten near the middle, where they are nearly coincident with the plane of maximum shear.

A device used by Mead[1] for making joints consists of a paraffin-coated sheet of rubber stretched on a rectangular frame of gas pipe. The paraffin can be subjected to compression, tension, torsion, or shear, or any combination of these stresses. Mead's results are as follows:

Non-rotational compression develops thrust faults striking at right angles to the axis of shortening and dipping about 45°; then vertical faults near the free edges of the sheet, at angles of 45° to the direction of shortening, due to the spread of the sheet; also a few vertical tension joints in the direction of the shortening.

Shear or rotational stress causes: (1) vertical tension cracks striking normal to the line of greatest elongation or 45° to the shearing movement; (2) two sets of vertical shear planes parallel to the sides of the distorted rubber sheet, or parallelogram; (3) thrust faults at right angles to tension faults (1) and dipping 45° either way.

In torsion, when the free edges of the rubber are stretched, tension cracks form along free margins and converge toward the center where they disappear. When the free edges maintain constant length and a line within the rubber midway between them is shortened, the results are similar to those in pure shortening.

In pure elongation vertical tension cracks are developed at right angles to the direction of elongation.

B. Folds

(*a*) *Strata folded by endwise compression.* The first experiments on folding were made by Hall[2] as an outgrowth of his

[1] Mead, W. J., Mechanics of geologic structures: Jour. Geol, vol. 28, 1920, p. 505.

[2] Hall, J., Vertical position and convolutions of certain strata and their relation with granite: Trans. Roy. Soc. Edinburgh, pt. 1, vol. 7, 1814, pp. 79–108.

studies of folds on the cliffs of Berwickshire. Hall believed that strata were folded by essentially horizontal thrust due to upwelling intrusives, mainly granitic masses. He, therefore, used endwise compression in his folding experiments.

Hall's apparatus was a box each of the end boards of which could be pushed horizontally by means of a screw. The lid was clamped down. Clay layers and strips of cloth were used in separate experiments. His sketches of the completed experiments show a series of rather uniform parallel folds without "piling up" near the push boards.

Daubrée [1] developed folds by compressing variously colored layers of wax. The apparatus used was an iron box with screws for applying horizontal and vertical pressure. The conditions and results of the experiments were as follows: (1) Layers of uniform thickness and composition under uniform vertical pressure, with horizontal compression slowly increased. Folds increased in number as experiment progressed. (2) A bed of uniform thickness and uneven load and increasing horizontal compression. Folds formed in region of least load. (3) Layer thinnest at one end. Thinnest portion of bed buckled first, whether near or far from thrust. (4) Layer thinnest in middle. Folds formed first where layer was weakest.

Pfaff [2] used an elongated box supplied with a pressure board at one end for experiments on folding. Clay layers buckled in this way showed the usual piling up in front of the pressure board common in experiments of this type. Even wedge-shaped layers, thickest near the pressure board, buckled first nearest the thrust.

The key idea of Reade's [3] experiments is that expansion of earth material due to rise of temperature is the cause of folding. Tests made by him on the expansion of metals, sand-

[1] Daubrée, A., Géologie Expérimentale, 1879. Taken from W. Paulcke, Das Experiment in der Geologie.

[2] Pfaff, F., Der mechanismus der Gebirgsbildung, 1880.

[3] Reade, T. M., The origin of mountain ranges considered experimentally, structurally, dynamically, and its relation to their geologic history, 1886.

stones, limestones, and marbles are inferred by him to support this idea.

Folds in clay layers produced by compressing the lower members of a series more than the upper were regarded by Reade as in better harmony with folds in nature than other types, his view being that the lower layers of the earth's crust have been subject to swelling and wrinkling through rise of temperature.

In connection with studies of Appalachian structures, Willis [1] made folds by compressing layers of beeswax. Those requiring competency were hardened with plaster of Paris. Weak layers were made by softening beeswax with turpentine. Competency and load were found to have inverse effects on folding. Folds of large cross-section developed when the competency was great as compared with the load. The pressure was not transmitted far. Folds piled up in front of the pressure board until resistance to upward movement was great enough to make the thrust effective in turning up new folds farther back. A strong initial dip of the beds, however, started folds at a distance from the pressure board.

For making folds, Reyer [2] used weak strata of mud, sand, viscous liquids, and dough-like materials. Strata of these materials were laid down in a thin box and made to flow and glide by tilting. Both folds and faults were formed in this way. Folding and faulting by subaqueous gliding were imitated by causing inclined mud layers covered by water to flow and glide under their own weight. His results are illustrated by numerous plates.

Paulcke [3] aimed to reproduce Jura eastern and western Alpine structures. The apparatus used was an iron box in which layers could be compressed endwise. Rectangular portions of the bottom of the box could be raised or lowered

[1] Willis, Bailey, The mechanics of Appalachian structure: 13th Ann. Rept., U. S. Geol. Survey, pt. 2, 1893, pp. 211–281.

[2] Reyer, E., Geologische und geographische experimente, Leipsig, 1892–1894, 4 Hefte.

[3] Paulcke, W., Das experiment in der geologie, 1912.

either before or during the experiment. The load could be distributed in various ways. The strata used were gypsum, sand, and clay.

Jura structure was imitated by using alternating layers of gypsum and clay which had an initial down-warp near the middle of the box. For west Alpine structure, the layers were upraised at the end of the box away from the piston and heavily loaded near the piston. Layers of unequal competence laid end to end in the compression box were used for the experiment on east Alpine structure. Pressure was applied on the side nearest the strongest layers.

Rinbach [1] made folds by endwise compression of layers of sand and clay laid in a box. No load was used. The push block differed from other devices of this type in being no higher than the strata before deformation. Rinbach's plan was to study the buckling of strata against rigid units or "massifs," the push board representing a "massif" lying with vertical contact against a stratified unit. A tin box with gently sloping front placed before the push board so as to dip under the layers represented a "massif" with surface extending under the strata for some distance. The layers were arranged in various ways with respect to competency.

In his experiments a fold started near the push block or massif, usually with a low-angle overthrust fault dipping toward the strata. The fold overturned and the under side rode forward over the surface of the push block. On the limb away from the push block minor folds and thrust faults dipping toward the push block were developed.

Rinbach's experiments also dealt with the arrangement of cleavage in deep-seated folds, the sinking of areas of sedimentation, and the association of igneous activity with folding.

Hungerer [2] caused a plate of rubber to be arched into an anticline by endwise pressure, the sides being free. The axis

[1] Rinbach, C., Versuche der Gebirgsbildung: Neues Jahrb., B. B., XXXV, 1913, p. 689.

[2] Hungerer, E., Ein Belegstück zur Elastizitäts Theorie der Faltung: Centralblatt f. Min., No. 4, 1922.

of the resulting anticline was saddle-shaped, lowest in the middle and highest at the ends. The same experiment with soft iron, tin, resin and wax, pitch and butter, gave similar results. He describes a small, natural anticline having the saddle form.

(*b*) *Layers folded by contraction of a rubber substratum.* Favre [1] placed a layer of clay 25 mm. thick on a stretched rubber band. The ends of the layer were held by wooden strips. No load was applied. On slowly contracting the rubber arches and troughs were formed at right angles to the axis of contraction. Shrinkage of the earth's nucleus, because of loss of heat, was regarded by Favre as the cause of folding.

Schardt [2] followed Favre's method of using a stretched rubber sheet as the contracting medium upon which layers of various consistencies and thicknesses were placed. Folds of various types and thrust faults were obtained. The mechanics of the processes were described and deductions made from the results. The results were compared with occurrences in nature. (Taken from Paulcke, Das Experiment in der Geologie).

By relaxing a sheet of rubber stretched over a hemisphere of wood, Meunier [3] obtained anticlinal bulges which encircled the pole. These he compared to the major European mountain ranges.

Rinbach [4] partially covered an inflated rubber balloon with a thin layer of wet sand. Slight deflation of the balloon developed a group of polygonal, commonly five and six-sided, fracture figures due to overthrusts. The cracks often formed a mesh-like design with crescent-shaped sides, recalling,

[1] Favre, A., Archives des Sciences Physiques et Naturelles, Nouv. Per Tome 62 (Geneve 1878).

[2] Schardt, H., Études géologiques sur le Pays — d'Enhaut Vaudois. troisième partie. A. mecanisme des dislocations: Bulletin de la Soc. Vandoise des Sciences Naturelles, vol. 20, chapitres 15–17, 1884.

[3] Meunier, S., La Géologie Expérimentale, Paris, 1899.

[4] Rinbach, Carl, Versuche über Gebirgsbildung: Neues Jahrb., B. B. XXXV, 1913, p. 689.

according to the author, certain tectonic lines of the Eastern and Western Antilles and of Central America. When a straight, inclined cut was made in the sand, contraction caused an overthrust. Extensions of the crack through shortening were oblique to the initial direction. Long bands of clay were laid on the balloon so that their ends were slightly separated. On contraction the ends of the bands approached each other and finally slid by each other. He also experimented with the effect of relatively rigid areas on the balloon's surface, simulating "massifs." On contraction the massifs were inert, while folds were piled up around their margins.

Mead's [1] experiments on folding involve the effect of both pure shortening and rotational stress or shear. The latter condition does not appear to have been studied in earlier experiments. Like Favre, he used a stretched rubber sheet for the base of the layers and the contracting medium. The stratum used was a mixture of beeswax and turpentine, covered with a thin sheet of tin foil or rubber to give it competency. For purposes of photographing, plaster of Paris reproductions of the deformed surface were made. With pure shortening a series of overlapping folds trending at right angles to the axis of shortening was formed. The folds pitch at each end and show tension cracks along their crests. With rotational stress folds like those resulting from pure shortening were produced parallel to the axis of elongation, and thus inclined to the direction of shear.

C. Rock Cleavage

Experiments on rock cleavage hinge mainly on the question whether cleavage is dependent on the parallelism of the constituent minerals. Disagreement arose over the fact that the term rock cleavage has been applied to both fracture and flow cleavage.

[1] Mead, W. J., Mechanics of geologic structures: Jour. Geol., vol. 28, 1920, pp. 505–523.

Sorby [1] produced cleavage by compressing a baked mixture of iron oxide and pipe clay. Under one-sided pressure the particles were rotated into axially parallel positions at right angles to the pressure.

Tyndall [2] produced a fracture cleavage by compressing wax, and concluded that pressure and not the position of mineral axes controls cleavage.

Daubrée [3] caused plastic micaceous clay to flow through a cylinder. A rude cleavage developed in this way was due to the mica plates aligning themselves parallel to the direction of flow.

From experiments on the compression of partially molten substances, Daubrée [4] concluded that cleavage could develop by pressure in viscous magmas.

Under non-rotational strain Adams [5] caused the elongation of calcite grains in marble at right angles to the direction of pressure, or parallel to the longest axis of the strain ellipsoid. Shear planes developed under the same conditions were diagonal to the direction of crystal elongation.

Becker and Day [6] have shown that although crystals can grow against pressure, their growth in a plane normal to the pressure is greater, the proportion being 1,000 to 1 or still larger. Their experiments were made with alum.

By subjecting blocks of ceressin to a simple rotational stress, Becker [7] produced a slaty or fracture cleavage parallel to the planes of no distortion. He argued that the major axes

[1] Sorby, H. C., Slaty cleavage as exhibited in the Devonian limestones of Devonshire: Phil. Mag., 4th ser., vol. 11, 1856, p. 29.

[2] Tyndall, J., Comparative view of the cleavage of crystals and slate rocks: Phil. Mag., vol. 12, 1856, p. 35.

[3] Daubrée, A., Géologie Expérimentale, vol. 1, 1860, p. 413.

[4] Daubrée, A., Études et expériences synthétiques sur le métamorphisme: Annales des mines, 5^{e} serie, 1859–1860. (Taken from W. Paulcke, Das Experiment in der Geologie, 1912.)

[5] Adams, F. D., and Nicholson, J. T., An experimental investigation into the flow of marble: Phil. Trans. Roy. Soc. of London, vol. 195, 1901.

[6] Becker, G. F., and Day, A. L., The linear force of growing crystals: Proc. Wash. Acad. Sci., vol. 7, 1905, pp. 283–288.

[7] Becker, G. F., Experiments on schistosity and slaty cleavage: Bull. 241, U. S. Geol. Survey, 1904.

of crystals which form under this type of stress are parallel to the fracture planes, but offered no experimental proof. This view is opposed by Leith, who holds, on inductive grounds, that crystal elongation is parallel to the longest axis of the strain ellipsoid.

Wright [1] subjected heated cubes of glass obtained by vitrifying wollastonite, diopside, and anorthite, to differential pressure. In most cases crystals grown under these conditions were elongated in the direction of greatest relief.

Tarr and von Engeln [2] subjected pond ice contained in an unyielding metal cylinder to a pressure of about 1400 pounds to the square inch. Under this condition the principal axes of the ice crystals turned at right angles to the pressure, involving a shift of 90°.

D. Strength Tests on Stone [3]

Strength tests on stone include resistance to crushing, transverse shearing, and tensile stresses. Crushing strength is the greatest. The others follow in the order named. The texture of the rock controls strength to a greater degree than the hardness of the individual minerals. Reports by Buckley,[4] Hirschwald,[5] and the Watertown Arsenal [6] contain data on strengths of various rocks. Data on the subject are also found in many other reports. Tests on resistance to impact are reported by Föppl [7] and Goldbeck.[8]

Hirschwald tabulated ratios between different strengths of

[1] Wright, F. E., Schistosity by crystallization: Am. Jour. Sci., vol. 22, 4th ser., 1906, pp. 224–230.

[2] Tarr, R. S., and von Engeln, O. D., Experimental studies of ice with reference to glacier structure and motion: Zeitschrift f. Gletscherkunde, vol. 9, 1915, p. 81.

[3] See also Chapter XIII.

[4] Buckley, E. R., Building and ornamental stones of Wisconsin: Bull. 4, Wis. Geol. and Nat. Hist. Survey, 1898.

[5] Hirschwald, J., Handbuch der bautechnischen Gesteinsprüfung, 1912.

[6] Tests on metals, Watertown Arsenal, Mass., 1894 and 1895.

[7] Föppl, A., Mitt. aus dem Mech.-Tech. Lab., München, Heft 30 and 32 1906 and 1912.

[8] Goldbeck, A. T., and Jackson, F. H., The physical testing of rock for road building: Bull. 44, Office of Public Roads, 1912.

various rocks but holds that no constant ratio exists. Average ratios of tensile to crushing strength of granite, sandstone, and limestone were 1:35, 1:34, and 1:16 respectively. The ratios of shearing (scission) strength to crushing strength for granite, sandstone, and limestone were 1:14, 1:12.9, and 1:12.1. Between transverse (bending) strength and crushing strength, the average ratios for granite, sandstone, and limestone are 1:14.4, 1:10.6, and 1:8.4.

Rinne[1] holds that figures ordinarily obtained for crushing strengths are too high because of the restraining influence of the jaws of the machine. Short columns therefore give higher results than long columns. In columns beyond a certain length, bending stresses greatly reduce the crushing strength. By substitution of a rubber pad between the pressure plate and the test piece for the jaws ordinarily used, he obtained nearly uniform results with columns of various lengths. By experimentation he also found an ideal length of column for Carrara marble, in which the effect of the jaws is eliminated but which is not long enough to bring bending stresses into play. Both circular and square columns of this ideal length had the same crushing strength. The fractures accompanying failure of Carrara marble had intersections of 60 and 120 degrees, the acute angle being bisected by the axis of compression.

E. Gliding and Twinning of Crystals Compressed along a Single Direction by Non-Rotational Stress

Enclosed crystals or rocks under non-rotational vertical stress tend to yield by shearing along the planes of no distortion of the strain ellipsoid and by slips along certain crystal planes of the mineral grains. The internal slips take place with or without twinning. Karman[2] found that under ordinary crush-

[1] Rinne, F.; Vergleichende Untersuchungen über die Methoden zur Bestimmung der Druckfestigkeit von Gesteinen: Neues Jahrb. f. Min. Bd. I, 107, p. 45.

[2] Karman, A., Festigkeitsversuche unter allseitigem Druck: Zeit. des ver. deutscher Ing., Nr. 42, Bd. 55, 1911, p. 1749.

ing tests very little secondary twinning is developed in marble, but that when the pressure is high on all sides, failure without fracture is accompanied by abundant twinning of the marble grains.

Similar observations are reported by Adams,[1] who also found that gliding was favored by high temperatures.

The position of the gliding planes and the qualitative conditions of their development have been studied for a large number of minerals. Mügge [2] found that gypsum, antimony, bismuth, halite, calcite, ice, galena, cyanite, fluorspar, apatite, anhydrite, orpiment, vivianite, lorandite, graphite, molybdenite, brucite, mica, sylvite, and diopside developed gliding planes when subjected to one-sided compression at room temperature, the sides of the test pieces being supported. The samples tested were embedded in lead and compressed by a piston or subjected to hammer blows.

Vernadsky [3] named seventy-seven minerals which have gliding planes, among them hornblende, topaz, dolomite, corundum, beryl, tourmaline, and epidote.

Veit [4] used Kick's method for determining the gliding planes of minerals. Crystals were imbedded with sulphur or alum in a hollow cylinder of steel or copper. Vertical pressure was brought to bear on the long axis of the cylinder by means of a hydraulic press. By this method Veit determined the gliding planes on fluorspar, sphalerite, rhodochrosite, smithsonite, barite, aragonite, anhydrite, corundum, and hematite. Quantitative estimates of the stress conditions were impossible.

[1] Adams, F. D., and Nicholson, J. T., An experimental investigation into the flow of marble: Phil. Trans. Roy. Soc. of London, vol. 195, 1901.

[2] Mügge, A., Uber künstliche zwillingsbildung durch Druck aus Antimon, Wismut, und Diopsid: Neues Jahrb. f. Min., I, 1886, p. 183.

[3] Vernadsky, W., Die Erscheinnugen der Gleitung bei Kristallinischen Körpern: Wissensch. Ann. der Universitat Moskau, Naturio abt. Heft 13, 1897.

[4] Veit, Kurt, Künstliche Schiebungen und Translationen in Mineralien: Neues Jahrb. B. B. XLV, 1921, pp. 121–148.

F. The Flow of Compressed Solids Through Narrow Openings

By compression Tresca [1] caused laterally confined materials — lead, aluminum, and ice — to flow through narrow apertures.

Spring [2] welded powders of various metals, sodium chloride, and potassium chloride by compression in closed tubes, and caused the welded materials to flow out through openings. His conclusion that the compressed materials were liquefied has been questioned by Hallock and others (see below).

Tammann [3] found that with increasing pressure the constant rate of flow of metals through apertures follows a descending temperature curve.

G. Deformation of Solids Unequally Compressed on all Sides by Non-Rotational Stress

Hallock [4] failed to check the work of Spring who claimed that solids could be liquefied under pressure. Hallock's method was to compress solids in a steel cylinder by means of a piston. Silver coins which he stuck in wax and paraffin remained intact after suffering a pressure of 6,000 atmospheres. Two double-pointed tacks imbedded in paraffin were similarly unaffected. Layers of antimony, beeswax, paraffin, lead, and ground bismuth, the latter in a paper roll, subjected to compression remained unchanged.

Kick [5] placed pieces of halite and Carrara marble in a copper shell and filled the remaining void with soft materials introduced in liquid form. Shellac, sulphur, and alum were used as the imbedding materials. Alum was found most satisfactory. The samples were subjected to vertical compression.

[1] Tresca, H., Mémoire sur l'écoulement des corps solides soumis a de fortes pressions: Compte. rend des séances de l'Acad. des Sc. 59, 1864, p. 754.

[2] Spring, W., La plasticité des corps solides et ses rapports avec formation des roches: Bull. de l'Académie de Belgique Cl. des Sciences III, 37, I.

[3] Tammann, G., Kristallisieren und Schmelzen. Ein Beitrag zur Lehre der Änderungen des Aggregats zustandes, 1903.

[4] Hallock, W., The flow of solids: Am. Jour. Sci., vol. 34, 1887, p. 277.

[5] Kick, F., Die Prinzipien der mechanischen Technologie und die Festigkeitslehre: Zweite Abh. Zeitschr. des ver. deutscher Ing., 1892, p. 919.

The resulting deformation of the test pieces took place without cracking. A design was stamped on the marble. No quantitative statement of the stresses involved could be made by this method.

Adams and Nicholson [1] deformed cylindrical pieces of marble enclosed in steel tubes by vertical compression. At ordinary temperatures the marble cylinders swelled in the middle, due to movement along faint shearing planes and internal gliding of the grains. At temperatures between 300° and 400° C. the columns were deformed by movement along crystal planes only. Water had no effect on the results obtained at high temperatures. The grains were flattened at right angles to the pressure.

Rinne [2] used Kick's method on calcite and marble. The test pieces were embedded in alum with a copper shell. The pressure applied, 12,000 kilograms per square centimeter, was believed to be equivalent to a depth of 5,000 meters. Both calcite and marble samples were "ironed out" under these conditions. Failure took place by granulation along shear planes and by internal gliding of the grains. Sylvite [3] and halite were deformed without fracture by this method at pressures of 2,000 to 3,000 kilograms per square centimeter.

Woolson [4] subjected a column of concrete encased in a steel jacket to vertical compression which resulted in bulging the concrete without fracture.

Adams [5] reports experiments on the flow of marble up to temperatures of 1000° C. At pressures of 120,000 to 130,000 pounds per square inch and ordinary temperature, distortion

[1] Adams, F. D., and Nicholson, J. T., An experimental investigation into the flow of marble: Phil. Trans. Roy. Soc. of London, vol. 195, 1901.

[2] Rinne, F., Beitrag zur Kenntniss der umformung von Kalkspat und marmor unter allseitigen Druck: Neues Jahrb., vol. 1, 1903.

[3] Rinne, F., Beitrag zur Kenntniss der umformung von Sylvin und Steinsalz unter allseitigen Druck: Neues Jahrb., I., 160, 1904.

[4] Woolson, H., Some remarkable tests indicating flow of concrete under pressure: Eng. News, vol. 54, 1905, p. 459.

[5] Adams, F. D., and Coker, E. G., An experimental investigation into the flow of rocks — The flow of marble: Am. Jour. Sci., vol. 29, 4th ser., 1910, p. 465.

took place by fracture of the grains and by internal gliding. After a period of rest the deformed samples showed increase in strength. At higher temperatures, the internal motion of the grains was easier, and no fracture developed. The presence of water had no effect on the results. The specific gravity remained unchanged. Slow compression was favorable to adjustment by internal gliding. A column compressed 64 days had double the strength of one compressed only ten minutes.

With Kick's method of embedding test pieces in alum or other soft materials within a copper shell, Adams [1] found that a series of minerals which have a hardness of 5 or under were deformed without fracture. Harder minerals developed gliding and twinning planes but no marked flow. Very hard minerals were broken into powder. Soft rocks behaved like soft minerals. Hard rocks crumbled. Granites assumed a faint gneissic texture by granulation and rotation of the grains.

Adams [2] also subjected a column of diabase jacketed with a steel case to vertical compression at temperatures of 450° C. Adjustment took place by granulation along shear planes.

Karman [3] compressed marble and sandstone columns placed in steel cylinders, and showed that the elastic limit increases with increasing pressure from all sides. The machine used enabled him to control lateral and vertical pressures independently. Deformation took place by granulation along shear zones and by slipping along twinning planes within the crystals. The latter was dominant when pressure on all sides was very high. In ordinary crushing tests twinning was not markedly increased.

The principal deductions made from a study of Karman's results are as follows:

[1] Adams, F. D., An experimental investigation into the action of differential pressure on certain minerals and rocks, employing the process suggested by Professor Kick: Jour. Geol., vol. 18, 1910, p. 489.

[2] Adams, F. D., Experimental investigation into the flow of diabase: Bull. Geol. Soc. Am., vol. 21, 1910, p. 773.

[3] Karman, Th. v., Festigkeitsversuche unter allseitigen Druck: Zeit. des Ver. deutscher Ing., Bd. 55, 1911, p. 1749.

Under all conditions of lateral pressure covered by the experiments, deformation is largely elastic at the start. With increase in lateral support, the range of stress difference during which deformation is largely elastic becomes greater.

The turning point from largely elastic to permanent deformation, or elastic limit, is sharp with low lateral pressures. Its position becomes more uncertain with high lateral pressures.

The elastic limit increases at a slower rate than the lateral pressure. In the case of marble its rise for lateral pressures greater than 1650 atmospheres becomes slight.

As the vertical applied pressure is increased, permanent deformation increases more rapidly than stress difference. With low lateral pressures this condition soon changes to one of continued deformation under decreasing stress differences. With medium lateral pressure, it tends to change to continued deformation under constant stress difference. With high lateral pressure, deformation gains very slowly over stress difference and gradually tends to become uniform for constant stress difference.

A study of the closure of cavities by vertical compression in rock cylinders which were restrained laterally by shrunk-on steel jackets was reported by Adams [1] in 1912. No estimates of the stress differences or of the lateral pressures were given. At ordinary temperatures, bore holes in Solenhofen limestone began to close when the vertical pressure was somewhere between 193,000 and 200,000 pounds per square inch. Under a vertical pressure of 128,000 pounds per square inch, a slight change in a transverse bore hole was noted after two and one-half months. At a temperature of 450° C. slight contraction of the bore holes was noted when the vertical pressure reached 96,000 pounds per square inch. A pressure of 64,000 pounds per square inch and this temperature brought about no change after seven hours.

Granite showed no change when under a pressure of 160,000

[1] Adams, F. D., An experimental contribution to the question of the zone of flow: Jour. Geol., vol. 20, 1912, pp. 97–118.

pounds per square inch for two and one-half months. The bore holes began to fill with rock powder split from the walls under vertical pressures of 200,000 and 222,500 pounds per square inch, each lasting for two and one-half months. Pressure of 96,000 pounds per square inch with temperatures of 450° to 555° C. brought about no change in granite.

Adams [1] subjected rocks enclosed in steel cylinders of .25 and .33 centimeters thickness, respectively, to vertical compression. The rocks used were granite, diabase, slate, Belgian black limestone, dolomite, marble, sandstone, alabaster, and steatite. He found that the stress differences which caused deformation under these conditions were much greater than under conditions of no lateral pressure. The tests were made under constantly increasing lateral pressure due to the action of the steel cylinders.

Bridgman [2] compressed sands of quartz, feldspar, and talc by a non-rotational stress of 30,000 kilograms per square centimeter without effecting fusion. After compression the sands had formed weakly coherent buttons showing a faintly lamellar structure. On resting, the lamellae swell and fall apart. The compressed material still had a considerable percentage of voids. The failure to form a coalescent mass, Bridgman believes, may have been due to air films covering the particles.

H. Deformation of Solids by Hydrostatic Pressure

Bridgman [3] made experiments to test the three theories of rupture under conditions of hydrostatic pressure. The theory that rupture takes place when either principal tensile stress exceeds a certain value was tested by subjecting steel and

[1] Adams, F. D., On the amount of internal friction developed in rocks during deformation and on the relative plasticity of different types of rocks: Jour. Geol., vol. 25, 1917, p. 597.

[2] Bridgman, P. W., Failure of cavities in crystals: Am. Jour. Sci., vol. 45, 4th ser., 1918, pp. 243–268.

[3] Bridgman, P. W., Breaking tests under hydrostatic pressure and conditions of rupture: Phil. Mag., vol. 24, 1912, p. 63.

glass bars to hydrostatic compression, the ends of the bars being free. The fracture of the bars took place between the packing cases, glass making a clean break across the bar. Steel broke by "pinching off" in the manner common to tensile breaks of this substance.

For testing the theory of maximum stress difference, hollow cylinders closed at each end were subjected to hydrostatic pressure over the entire surface. When the inner tubes were perfect in bore, they became closed without rupture. The elastic limit was greatly increased. Yield took place but no rupture. With copper the yield point was raised ten-fold. The microscope showed elongation of grains in direction of flow. Pressures applied successively showed an increase in yield point each time. For equal increments of yield point, the diameter of the cylinders showed the same decrease. Glass tubes were subjected to a compression of 24,000 atmospheres without breaking. The maximum stress difference is on the inner surface of the tubes, and for a thick tube is equal to the hydrostatic pressure. The tensile strength of the glass was only about 7,000 atmospheres. Under the conditions of the experiment rupture took place when a stress difference fifty times this was reached.

The third theory of rupture, that a break takes place when the extension in any direction exceeds a critical value, was tested by subjecting cylinders of steel and glass to internal hydrostatic pressure. On the theory of principal stress, rupture should take place when the internal pressure equals the tensile strength. The theory of elongation, according to Bridgman, requires Bessemer steel to break when the internal pressure is equal to four-fifths of the tensile strength. With metals, rupture started on the outside along axial planes or along spiral shear planes. The inner surfaces were greatly elongated before rupture started. Glass capillary tubes should theoretically stand no more than 500 atmospheres, but some actually stood 1000 atmospheres internal pressure before breaking. With glass the break started on the inside.

Bridgman concludes that the three theories of rupture alluded to do not hold under the conditions of hydrostatic pressure outlined. Brittle and ductile substances behaved differently, however.

Bridgman[1] concluded that stresses twenty times greater than the crushing strength may be necessary to close cavities in rocks by hydrostatic pressure. The estimates by Adams of the depths necessary to close cavities he believes are too high, however, owing to the unknown action of shrunk-on steel jackets which affected Adams' results.

Bridgman used quartz, tourmaline, calcite, feldspar, andesite, porphyry, limestone, granite, and glass in his experiments. The test pieces were turned into cylinders and then sawed in two. The severed surfaces were polished and a nearly round hole was bored from one or both surfaces nearly coaxial with the cylinder. The two pieces were then fitted together in their natural position and held together by a cover of rubber tubing clipped at each end.

Hydrostatic pressure up to 12,000 kilograms per square centimeter was applied on all sides. Up to a certain pressure, varying with the rock type, no change took place. Beyond a certain pressure particles were broken from the walls of the cavity. In the case of crystals, the "flaking off" was not related to crystal directions. Cracks developed but no slipping took place. Only barite and granite showed changes in dimensions, indicating viscous flow according to Bridgman.

I. Effect of High Temperatures on Deformation of Solids

The belief that heat favors deformation of solids without rupture seems to be consistently upheld by experiments.

Milch[2] found that crystals of halite, melting point 800° C., were easily deformed under pressure at a temperature of 200° C.

[1] Bridgman, P. W., Failure of cavities in crystals and rocks: Am. Jour. Sci., vol. 45, 4th ser., 1918, pp. 243–268.

[2] Milch, L., Über Zunnahme der Plastizitat bei Kristallen durch erhöhung der Temperatur: Neues Jahrb., 1, 1909.

Nagaoka[1] found that high temperatures lower the modulus of elasticity of rocks, i.e., rocks become more compressible at high temperatures.

Day and Allen[2] observed that albite crystals when near the melting point were easily bent under a slight load.

Adams[3] in his experiments on the flow of marble and other rocks used temperatures up to 1,000° C., and found that deformation without shearing planes was greatly favored by heat and that the strength of marble deformed under conditions of high temperature was greater than that of marble deformed at lower temperature but under similar pressure conditions. The gliding phenomena of marble grains were more in evidence as a result of deformation at high temperatures.

J. Volume Changes of Minerals and Rocks Caused by Changes in Temperature and Pressure

(1) *Expansion and contraction of rocks through temperature changes.* The rates at which solids expand with rise of temperature is reported in terms of the coefficients of linear and cubical expansion. The coefficient of linear expansion is the ratio of the elongation of the substance caused by a rise of 1° C. to its original length. The coefficient of cubical expansion is the increase in volume of the substance when its temperature is raised 1° C., divided by its original volume. The coefficient of cubical expansion is about three times as large as the coefficient of linear expansion.

The coefficients of expansion of minerals and rocks are not the same for all ranges of temperature. In crystals other than isometric, they also vary with the crystal direction. Only

[1] Nagaoka, H., Modulus of elasticity of rocks and velocities of seismic waves: Publications of the Earthquake Committee, No. 17, Tokyo, 1904.

[2] Day, A. L., and Allen, E. T., The isomorphism and thermal properties of the feldspars: Pub. Carnegie Inst., Wash., 1905.

[3] Adams, F. D., and Coker, E. G., An experimental investigation into the flow of rocks — The flow of marble: Am. Jour. Sci., vol. 29, 4th ser., 1910, p. 465.

Adams, F. D., An experimental investigation into the action of differential pressure on certain minerals and rocks, employing the process suggested by Professor Kick: Jour. Geol., vol. 18, 1910, p. 489.

like directions crystallographically have the same coefficients of expansion.

Below are a few coefficients of expansion of common rocks listed in the Physico-Chemical tables of Castell-Evans. Additional data are given in the Tabbelen of Landolt and Börnstein. No. 5 is from the work of Barus.[1] A and B were adapted from the coefficients given. A represents the linear expansion in miles of 100 miles of the substance when its temperature is raised 1,000° C. B is the expansion in cubic miles of 100 cubic miles caused by a rise of 1,000° C.

Material	*Temperature range.*	*Coefficient of linear expansion*	*Coefficient of cubical expansion*	*A.*	*B.*
1. Granite	0 – 100°	.00000868	.00002608	.868	2.61
2. Limestone	0 – 100°	.00000251	.00000653	.251	.65
3. Marble	0 – 100°	.00000455	.00001365	.455	1.36
4. Quartz	19 – 46°	.0000119	.0000357	1.19	3.57
5. Diabase	0 –1000°	....	.0000250	....	2.50

A. Linear expansion in miles of 100 miles caused by raising the temperature 1,000° C.

B. Expansion in cubic miles of 100 cubic miles when the temperature is raised 1,000° C.

Both A and B are based on the assumption of uniform expansion through a temperature range of 1,000° C.

(2) *Changes of crystal structure through temperature changes.* Crystals undergo changes in optical properties and in their interfacial angles when the temperature is raised. Some minerals are also known to change their crystallization at certain temperatures. Quartz has such an inversion point at about 800° C., as shown by Fenner [2] and others. At this temperature it changes into tridymite. At 1470° C. tridymite changes to cristobalite. Aragonite changes at 445° C. to calcite. For both silica and lime carbonate, the inversion forms which develop at the higher temperatures have the larger volume. The influence of pressure on inversion does not appear to have been studied by experiment.

[1] Barus, C., High temperature work in igneous fusion and ebullition, chiefly in relation to pressure: Bull. 103, U. S. Geol. Survey, 1893.

[2] Fenner, C. M., The stability relations of the silica minerals: Am. Jour. Sci., 4th ser., vol. 36, 1913, p. 331.

(3) *Chemical reactions as a cause of volume changes.* Temperature and pressure influence chemical reactions and volume changes are effected thereby. No attempt is made here to list the experiments which bear on this complex subject. Johnston, Niggli,[1] and Adams[2] have reviewed the work of Spring, Tammann, Spezia,[3] Masing,[4] and others in the influence which temperature and pressure have on chemical reactions. They also give results of their own.

Johnston and Adams conclude that uniform pressure promotes those reactions which are accompanied by decrease in volume, but that it has not been shown to be the cause of such reactions. Unequal pressure, they say, is more effective than uniform pressure in promoting reactions, diffusion, and solution.

(4) *Volume changes through change of phase.* Rocks occupy larger volume at atmospheric pressures in the liquid than in the solid phase. This has been shown by the work of Barus,[5] Douglas,[6] and others.

Barus showed that molten diabase contracts about 3 per cent in volume when it solidifies into the solid glassy state; and that the change from diabase to glass involves an expansion in volume of about 11 per cent. Douglas measured the volume changes involved in the change from the crystalline to the glassy state of fifteen different rocks. Granite showed more expansion than the other rocks, its volume increase being 10 per cent. The expansion of gabbro was 6 per cent. The basic rocks showed less expansion than the acid varieties. Douglas

[1] Johnston, J., and Niggli, P., The general principles underlying metamorphic processes: Jour. Geol., vol. 21, 1913, pp. 481 and 588.

[2] Johnston, J., and Adams, L. H., On the effect of high pressure on the physical and chemical behavior of solids: Am. Jour. Sci., vol. 35, 4th ser., 1913.

[3] Spezia, G., Some presumed chemical and physical effects of uniform pressure: Atti. R. Accad. Sci. Torino, XLV, 1910, pp. 1–16; Rivesta Min. Ital. XXXV, 1908, pp. 62–64; Atti. Accad. Sci. Torino, XLVI, 1, 1911.

[4] Masing, G., Zs. Anorg. Chem. LXII, 1909, pp. 265–309.

[5] *Op. cit.*

[6] Douglas, J. A., On changes of physical constants which take place in certain minerals and igneous rocks on the passage from the crystalline to the glassy state: Quart. Jour. Geol. Soc., vol. 63, 1907, p. 145.

also summarizes the results of Delesse, Bischof, Forbes, and Barus. The earlier experimenters used less refined methods and usually obtained somewhat larger figures for these volume changes than Douglas and Barus.

Rock is not the only product of the solidification of magma. Great quantities of gas are also set free. The gas pressure is therefore increased many times when the magma freezes. This is shown by Morley.[1]

Tammann [2] inferred from experiments that uniform pressure beyond a critical point lowers the melting point. This implies that, beyond the critical pressure, the volume of the substance is less in the liquid than in the solid state.

Work by Bridgman [3] did not confirm Tammann's results. Bridgman studied the effect of uniform pressure on the melting points of twenty different substances, at pressures ranging from 0 to 13,000 kilograms per square centimeter. The temperatures ranged from 0 to 200° C. He concluded that his results gave no indication that the melting point would be lowered by very high uniform pressure.

(5) *The effect of pressure on specific gravity. Permanent volume changes through pressure.* The results of experiments on the effect which pressure has on the specific gravity of substances seem to vary. In certain experiments hydrostatic pressure showed no effect on specific gravity. Some experiments with unequal pressure caused a decrease in specific gravity, while others caused no change.

Kahlbaum [4] found that certain metals subjected to unequal compression, with a maximum stress of more than 10,000 atmospheres, showed an increase in specific gravity. He believed that he used hydrostatic pressure but this view is held to be erroneous by Bridgman.

[1] Morley, G. W., The development of pressure in magmas as a result of crystallization: Jour. Wash. Acad. Sci., vol. 12, 1922, pp. 219–230.

[2] Tammann, G., Kristallisieren und Schmelzen, 1903.

[3] Bridgman, P. W., Changes of phase under pressure: Phys. Rev., N. S., vol. 3, Nos. 2 and 3, 1914; also Nos. 2 and 3, 1915.

[4] Kahlbaum, W. A., Roth, K., and Siedler, P., Uber Metall. distillation und über distillierte Metalle: Zeitschr. Anor. Chem., vol. 29, 1902, p. 254.

Adams [1] found no change in the specific gravity of marbles deformed without rupture under conditions of unequal pressure.

Bridgman [2] subjected metals to a hydrostatic compression of 25,000 to 30,000 atmospheres without effecting change in specific gravity.

Johnston and Adams [3] conclude that when substances are deformed a decrease in specific gravity takes place.

Lea and Thomas [4] report that steel columns which they deformed by unequal pressure showed a decrease in specific gravity. After a rest of thirty-five days, the specific gravity had increased again.

(6) *The elastic compressibility of rocks.* Rocks which are compressed within the limits of elasticity resume their original shape when the pressure is removed. In practice, rocks are not completely elastic, but show some permanent deformation when compressed. Porous rocks show more permanent deformation than compact varieties.

The elasticity of rocks, when subjected to pressure in one direction, the sides being unsupported, is reported in terms of Young's modulus of elasticity and Poisson's ratio. Young's modulus is the quotient of the longitudinal stress by the longitudinal shortening. Poisson's ratio is the lateral extension divided by the longitudinal shortening. These constants have been determined by various investigators for a variety of rocks.

Nagaoka [5] has determined the modulus of elasticity for rocks of various ages. In a later work,[6] he gives a few pre-

[1] Adams, F. D., and Coker, E. G., An experimental investigation into the flow of rocks — The flow of marble: Am. Jour. Sci., vol. 29, 4th ser., 1910, p. 465.

[2] Bridgman, P. W., Breaking tests under hydrostatic pressure and conditions of rupture: Phil. Mag., vol. 24, 1912, p. 63.

[3] Johnston, J., and Adams, L. H., On the effect of high pressures on physical and chemical behavior of solids: Am. Jour. Sci., vol. 35, 4th ser., 1913.

[4] Lea, F. C., and Thomas, W. N., Change in density of mild steel strained by compression beyond the yield point: Engineering, vol. 100, 1915, pp. 1–3.

[5] Nagaoka, H., Elastic constants of rocks and the velocity of seismic waves: Phil. Mag., vol. 50, 1900, p. 53.

[6] Nagaoka, H., Modulus of elasticity of rocks and velocities of seismic waves: Publications of the Earthquake Committee, No. 17, Toyko, 1904.

liminary results showing the effect of high temperature on elasticity. He finds that the modulus of elasticity is lowered by increase in temperature.

Adams and Coker[1] have also determined Young's modulus and Poisson's ratio for a variety of rocks. From these data they have computed what they believe to be the modulus of cubic compressibility of the rocks tested. They have then drawn certain inferences regarding the state of compression of rock materials in the depths of the earth. The modulus of cubic compressibility is the quotient obtained by dividing the uniform or hydrostatic pressure in pounds per square inch by the resulting diminution in volume per cubic inch.

From the moduli of cubic compressibility given by Adams and Coker, the percentage decrease in volume for a pressure of one pound against each side of an inch cube of various rocks is inferred to be as follows:

Black Belgian marble	.000012
Carrara marble	.000018
Baneno granite	.000022
New Glasgow gabbro	.000010
Sudbury diabase	.000009
Ohio sandstone	.000055

Work by Williamson[2] has shown that moduli of cubic compressibility computed from data obtained by compressing rocks by one-sided pressure, leaving the sides unsupported, cannot be taken as reflecting the elastic properties of rocks under uniform pressure. Williamson finds that under high hydrostatic pressures the compressibility of substances is less than at low pressures. He states that at 2,000 megabars[3] the compressibility per megabar, expressed in parts per 10,000,000, is as follows for each of the succeeding substances:

[1] Adams, F. D., and Coker, E. G., An investigation into the elastic constants of rocks, more especially with reference to cubic compressibility: Carnegie Inst., Wash., 1906.

[2] Williamson, E. D., Change of physical properties of materials with pressure: Geophysical Lab., Wash., No. 446, 1922.

[3] A megabar equals .987 atmospheres.

Marble	1.41
Granite	2.13
Basalt	1.88
Diabase	1.26

At a pressure of 10,000 megabars, the compressibility per megabar of the preceding substances, expressed in the same units as before, is:

Marble	1.41
Granite	1.84
Basalt	1.55
Diabase	1.26

The percentage of volume decrease for a cubic compression of one pound per square inch, when the pressure is 2,000 megabars, or about 30,000 pounds per square inch, is about as follows:

Marble	.0000009%
Granite	.0000014%
Basalt	.0000012%
Diabase	.0000008%

At a pressure of 10,000 megabars (about 150,000 pounds per square inch) the percentage decrease in volume for a cubic compression of one pound per square inch is about as follows:

Marble	.0000009%
Granite	.0000012%
Basalt	.0000012%
Diabase	.0000008%

K. The Rigidity of the Earth

Inferences as to the rigidity of the earth as a whole have been made from studies with the horizontal pendulum. With it the changes in direction of the gravitational vertical, due to the attractions of the sun and moon, are measured. Observations

of this type have been made by G. H. and H. Darwin,[1] Schweydar,[2] and others. The Darwins arrived at no definite conclusion.

Michelson[3] says that Schweydar, and others following Darwin, agree in a general way that the earth is about as rigid as steel.

The direction of the gravitational vertical has been studied by Michelson[4] by means of observing the fluctuations of the water level in two horizontal six-inch pipes, each 502 feet in length and buried six feet in the ground. One is laid east and west; the other north and south. Michelson infers from his studies that the rigidity of the earth may be greater than that of steel.

[1] Darwin, G. H., and H., On an instrument for detecting small changes in the direction of the force of gravity: British A. A. S., York meeting, 1881, pp. 93–126. See also 1882.

[2] Schweydar, W., Untersuchungen über die Gezeiten der Festen Erde. Potsdam, 1912. (Not read.)

[3] Michelson, A. A., Rigidity of the earth: Jour. Geol., vol. 22, 1914, pp. 96–130.

Michelson, A. A., and Gale, H. G., The rigidity of the earth: Jour. Geol., vol. 27, 1919, pp. 585–601.

[4] *Op. cit.*

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